ИЗВЕСТИЯ
АКАДЕМИИ НАУК СССР
СЕРИЯ ГЕОЛОГИЧЕСКАЯ

IZVESTIYA AKAD. NAUK SSSR

SERIYA GEOLOGICHESKAYA

1959 | No. 1, January

CONTENTS

BRIEF COMMUNICATIONS

REVIEWS AND DISCUSSIONS

T-13128. Approved for printing December 19, 1958
Circulation -- 4,200 copies. Order 3311
Paper size 70 x 108-1/16. Paper 4. Printing sheets 10.96. Publ. sheets 11.4

Second printing office of the USSR Academy of Sciences Publishing House.
Moscow, Shubinskiy per. 10.

GEOLOGIC SCIENCE AT THE TIME OF THE XXI CONGRESS OF THE COMMUNIST PARTY OF THE U.S.S.R.[1]

by

D. I. Shcherbakov

The attention of the Soviet people today is focused on the figures in the report by comrade N. S. Khrushchev at the XXI Communist Party Congress, on the 1959-1965 economic development of the U. S. S. R. The report outlines with utmost clarity the enormous problems facing the Soviet people: "Our country now enters a new and most important period of its development, the period of intensive building of a communist society. Our main problem is the establishment of an all-embracing, material and technical basis of communism, of further strengthening the economic and defensive might of our fatherland, and of providing a greater satisfaction of the ever-growing material and spiritual needs of the Soviet people."

After reviewing the amazing industrial and cultural achievements in the U. S. S. R. in the past years, the report outlines the plans for the future: "The main task of the 1959-1965 seven-year plan is the all-embracing development of the Soviet economy on the basis of preferential growth of heavy industry, to strengthen the economic potential of the country in order to insure a continuous rise in its standard of living."

It is emphasized that the next seven years will witness a more intensive utilization of the rich natural resources of our land, a more effective distribution of its productive forces, a closer contact of industry with the sources of raw material, fuel, and the consumption centers. Especial attention is given to progressive assimilation of the natural resources of the eastern regions of the U. S. S. R.

With this in mind, much emphasis is put on the development of the mineral resources of the country by means of an intensification and more rational organization of geologic prospecting. The overall volume of such work is to be increased by about 65 percent, chiefly in the field of oil, gas, and rich and easily developed ferrous and nonferrous ore deposits favorably located in relation to the planned locations of new industries.

The report also foresees a rise in the efficiency of geologic prospecting by using the best available methods, and by using new geophysical and drilling techniques.

Soviet science, with its new achievements and its industrial applications, is destined to play a leading part in the progress of our country toward communism. Science has become a great modifying force whose progress is the basis of the entire planned socialist economy. At the same time, the combined principles of planning, coordination, and interdependence of scientific effort are becoming a characteristic feature of Soviet science. This is definitely reflected in the planning of basic scientific research for the next seven years.

The Academy of Sciences, U. S. S. R., in cooperation with the academies of the union republics, the Ministry of Higher Education, and the State Scientific-Technical Committee of the Council of Ministers of the U. S. S. R., has outlined the directions to be followed, on the basis of a study of current scientific trends, both in the U. S. S. R. and abroad. This made it possible to chart the branches of science that are developing most rapidly and, as a result, will most broaden and deepen out knowledge of nature and of man in it. Following through in these directions will, moreover, have a great effect on the progress of industry, transportation, and agriculture. Such scientific study will contribute to the solution of the momentous task set forth for the people by the Special XXI Congress of the Communist Party, U. S. S. R.

The problem of broadening the mineral base of our economy is closely related to the achievements of geology, the main purpose of which is the correct evaluation of

[1]Geologicheskaya nauka k XXXI S'yezdu KPSS.

natural resources and their rational development, as well as a geologic and metallogenic zonation of the country. The diversity of industrial activity calls for the development of the various, closely interrelated branches of geology presenting, so to say, a single geologic front.

Diversification is the main feature of modern geology, which, since the turn of the century, has branched out in various directions with no less than 15 to 20 such branches, each branch going its individual way, with its own objectives and methods of achievement. Among these branches, there are several with a more important scope. They develop at a faster rate with the promise of greater "yield."

Foremost among the geologic disciplines serving the daily needs of mining and construction, and thus fulfilling the functions of "scientific service," are the fields of geochronology and stratigraphy, lithology, petrography and mineralogy, seismic zonation, volcanology, and volcanic zonation, Quaternary geology, and such applied geologic disciplines as hydrogeology, engineering geology, geology of coal and oil, etc. All are in a state of continuous development, and without them neither the compilation of modern geologic maps -- both general and detailed -- nor the solution of the geologic aspects of major scientific and economic problems is possible. This is why all these branches of geologic science must be developed over and above their application to any specific problem, with the rate of development controlled by the increase in demand in a particular field.

Of great importance is geochronologic and stratigraphic research, for it is the main link in the chain of geologic studies. Indeed, the foremost objective of geology as an historic science is the unravelment of the time sequence of events. It provides geologic research with the tool of correlation. Accordingly, both scientific and applied research clamor for geochronologic scales built upon biostratigraphic principles; also for the absolute age scales of various geologic formations, based upon study of terminal isotopes of natural radioactive elements. These studies should be carried on by all means, in the immediate future, because of great demand from oil and coal industries as well as from nonferrous and rare-metal industries.

Seismology and volcanology stand out among the branches of the geologic scientific service. Seismic service is now organized in nearly every country; where seismic prospecting is carried on, anti-earthquake construction is practiced in the provinces subject to earthquakes. Very interesting data are being collected, shedding light on the nature of seismic phenomena.

Of no less importance is the systematic study of active and recently extinct volcanoes. This study leads to valuable conclusions as to the causes of eruptions, the distribution of internal heat about the magmatic hearths and, consequently, to charting of comprehensive research in the field of geothermy. For immediate results, this would lead to identification and assimilation of a new kind of mineral resource, namely superheated plutonic waters.

Especially rapid has been the development in the study of sedimentary rocks, or lithology, as the science of uniformity in the formation of sedimentary rocks and associated mineral products. It deals with the laws of rock genesis and it is related to tectonics in so far as it considers the time and nature of the earth-crust oscillations, which are reflected in the composition of the sediments. It also considers the lateral distribution of matter, and it widely uses mineralogic, geochemical, petrographic, and stratigraphic methods, as well as experimental study.

The ever-growing importance of lithology is also explained by the fact that it regards the sedimentary rocks as a source of various chemical elements and it reveals hitherto unknown types of mineral deposits. In connection with the most recent data on depth circulation of hot waters, sedimentary rocks may also be regarded as a possible source of various metals and chemical elements which are extracted from them by these waters and then deposited, in one way or another, as a new source to be exploited.

Despite this obvious modern-geology differentiation into specialized branches, the opposite tendency is also clearly defined: that of complex problems to be solved by a joint effort of many sister disciplines. The time has come when new trends appear in the borderland of two or more disciplines, tending to tie them together. Thus, geophysics and geochemistry have been born in the last few decades, and now cosmogeology is in its inception.

Among the most important current complex problems is the determination of distribution characteristics of major useful minerals throughout the earth's crust, as the basis of prediction of their presence. This research direction, proposed by Academician N.S. Shatskiy, is gaining importance in modern geology. To be sure, not all geologists follow N.S. Shatskiy in his emphasis on distribution characteristics alone. Many

believe that the problem of ore-deposit origin is of no less importance. However, much discussion of this topic has led scientists to a quite correct conclusion that distribution patterns do exist with all their practical implications. As to the problems of origin, they are still highly speculative, rarely solved, and therefore not a good basis for practical solutions. Indeed, the two cannot be separated: in dealing with the characteristics of distribution, we dwell on the problem of ore-deposit origin which cannot be overlooked in prospecting.

The knowledge of the laws of distribution of mineral fuels, ferrous, nonferrous and rare metals, rare and scattered elements, fertilizers, chemical raw materials and various nonmetallic minerals will stand in good stead in their search, rendering this search more rational and economical.

The experience of determining the distribution patterns of useful minerals has shown that the exploration for them is best grouped as follows: 1) economic minerals of sedimentary origin; 2) those of magmatic origin; 3) oil and combustible gases; 4) coal and carbonaceous shales; 5) rare and scattered elements.

It is further evident that the ultimate result of such study should be reports and monographs on the subject, together with special metallogenic and exploratory maps. It is not true, as some investigators believe, that exploration maps alone are the main and most practical goal. It is well known that maps, by their very nature, cannot reflect all the characteristics of three-dimensional space, in addition to the time factor.

The determination of the distribution of oil and gas has high priority. As is well known, world production of oil and gas reached a very high level in 1957. Oil production has risen sharply also in our country, with the rate of growth maintained by the discovery of new oil and gas provinces and deposits.

The 1965 annual production of oil is planned at 240 million tons (metric), and 150 billion cubic meters of natural gas. Such problems are to be solved by broadening the mineral base and by creation of ever-increasing proven reserves of oil and gas. Therefore, discovery of new oil and gas provinces is called for, along with the increase in proven reserves in the old ones, on the part of our scientific research and exploration organizations.

It is urgent to scientifically evaluate the potential of vast regions of Siberia, Kazakhstan, Central Asia, and of other provinces of the Soviet Union. Geology is to play a proper part in this major, state-wide task. The problems of the development of oil and gas production must be solved on the basis of a comprehensive analysis of the geologic structure of our land and of individual regions.

Exploration in general, and especially the compilation of exploration maps is greatly assisted by the study of tectonics, the science of structure and development of the earth's crust, in a dynamic stage. The voluminous data in this field first require organization and generalization. The best and most graphic means of such generalization are tectonic maps, i.e., maps showing the structural forms of the earth's crust, with the distribution of its component elements according to their age.

Another scientific task, also closely related to the broadening of the mineral base, is the development of prospecting methods for so-called "blind and covered" ore bodies, i.e., those not exposed and consequently not subject to visual observation. This is true for both new and unexplored areas and for known ore deposits.

Most diversified characteristics in the structure and development of the earth's crust influence the search for hidden ore deposits. The factors and their relationship to be considered are structural forms, sedimentation, magmatism, and weathering phenomena. The time of the formation of this or that ore deposit must be considered and proper search methods applied for each kind.

However, working in new provinces calls for immense capital expenditures in the organization of new production, especially if it is found in poorly accessible and sparsely populated regions. It is natural, then, for the exploration geologist to turn first to the known mines and mine fields. There always is the legitimate question as to whether the ore reserves of a given mine cannot be enlarged, both laterally and in depth. A very respectable and theoretically feasible task, is then, the search for hidden ore bodies in the area of active mines.

This will require a structural study of mine fields together with the study of the principles of forecasting at depths, on the basis of paragenetic associations of minerals, or the study of the formation of such associations. Other exploration methods are related to the geochemistry of so-called "halos of scattering." In any event, the detailed geologic position of ore bodies must be determined. Especially important is the so-called "structural control," i.e., the

development characteristics of a fracture system controlling the movement of mineralizing solutions of their opening and closing, as well as the effect of the enclosing rock.

The theory of vertical zonation is also of great importance in the forecasting of the depth of hidden ore bodies. The term, vertical zonation, means the regular alternation of mineral associations with depth, observed in all ores deposited from hot aqueous solutions. In explaining this vertical zonation and the alternation of mineral associations, the leading part should be assigned to the replacement, or metasomatic, processes widely developed in nature. In regarding these processes as consecutive stages of the action of solutions upon the solid body of ore veins and the emplacing rocks, they may be made subject to the laws of thermodynamics and to the rule of phases. Thus, with the nature of a solution determined, its changes with depth may be forecast on a physical-chemical basis.

Geochemistry, which deals with chemical processes in the earth's crust as well as with its chemical composition, opens new horizons with regard to the origin of ore deposits.

Modern geochemistry has branched out in three directions. First, it is statistics of the precise chemical composition of the earth's crust. Second, it is crystallochemistry, which has become an autonomous branch dealing with the chemical composition characteristics of crystalline substances. Finally, it is isotope analysis, often called isotope geology. A precise knowledge of the differentiation characteristics of isotopes in minerals and rocks makes it possible to solve some of the major and more general problems in geology. Such modern methods as mass spectroscopy provide valuable data for a more detailed approach to the problems of the origin of ore deposits, sea water, atmosphere, etc.

A development of modern trends in geology assures more rapid, more comprehensive, and cheaper methods of utilizing the natural wealth for the service of mankind.

Especially fruitful is the new scientific research planning, following new trends along with new major scientific goals. This marks a new stage in the organization of scientific effort, with emphasis on the more important sectors of the scientific front.

As forcefully expressed in the Report: "A pledge of a successful fulfillment of the great task of further development of socialist construction, planned for 1959-1965, is the self-denying effort and creative initiative of the heroic toilers' class, of our glorious scientific and technical intellectuals, and of all the Soviet people."

THE PROBLEM OF THE TECTONIC POSITION OF THE RIPHEAN VOLCANICS ON THE RUSSIAN PLATFORM[1]

by

A. S. Novikova

The development of tectonic forms in the sedimentary mantle of the Russian platform began with the formation of narrow troughs, breaking the crystalline foundation into several large blocks, with a series of small crystalline ridges formed along the eastern and southwestern rim of the platform, separated by deep depressions filled with sediments.

The Riphean volcanics are best developed along the edges of the Russian platform and on the flanks of the troughlike downwarps.

* * * * *

INTRODUCTORY REMARKS

A prominent part in the Riphean deposits of the Russian platform is played by volcanics found in the southwestern Belorussia, Kresttsy, Kaluga region, and in other localities. Volcanic material occurs either scattered in various admixtures with clastic rocks (Nepeytsino, Minsk) or in tuffaceous beds and sequences of beds (southwestern Belorussia, Kaluga region, Kresttsy). Lava flows, chiefly basic, are found at places (Kaluga region; Kaltasy-Bashkiriya; Staro-Petrovo; Kresttsy; Sporovo, southwestern Belorussia; Ivatsevichi, Malorita, Ivanovo, Manevicha, and other localities). Deep drilling revealed gabbroid vein rocks cutting the assorted crystalline basement rocks of the Russian platform, in a number of places. These formations are found in southwestern Belorussia, eastern Tatariya, and western Bashkiriya.

The purpose of this paper is to clarify the tectonic position of basic volcanics among the Riphean deposits. The topic of the regularity of volcanic manifestations on the Russian platform has never been touched upon, and this paper is the first such attempt.

The author is pleased to express her gratitude to N. S. Shatskiy for numerous consultations on the geology of the Russian platform.

* * *

In the western and central parts of the Russian platform, ancient deposits rest underneath faunally determined Lower Cambrian formations. In the eastern and southern parts, they are dovered by the Devonian. Most students are in agreement that formations correlative with those underlying the Lower Cambrian of the western and central parts of the platform are developed among those underlying the Devonian.

N. S. Shatskiy [31] divides this ancient section into three parts. The lower part is made up of the Kaverino terrigenous sequence; the middle part, of terrigenous-dolomitic rocks bearing local names in different areas. Here belong the Orsha, Pinsk, Nenoksa, and Bavly formations, as well as the Redkino and Serdobsk sequences. The terrigenous-dolomitic formations are followed by areno-argillaceous rocks, the Valday of the north and west of the platform, known as the upper Bavly formations in the east, where it is correlative with rocks overlying the Serdobsk sequence of the Pachelma downwarp. On the index map (Fig. 1), the lower part of the Riphean, corresponding to the Kaverino sequence, is designated by the symbol 'K.' Rocks of the Orsha, Pinsk, Nenoksa and lower Bavly sequences are marked with the first letter of their name. The overlying Valday, upper Bavly, and upper and lower Pachelma rocks are designated Вк, Вбс, Bпс, Нпс, respectively.

The age of these ancient deposits is under discussion. B. S. Sokolov [19] regards the Valday (Vendian) deposits as Sinian, belonging

[1] K voprosu o tektonicheskom polozhenii rifeyskikh vulkanogennykh porod na Russkoy platforme.

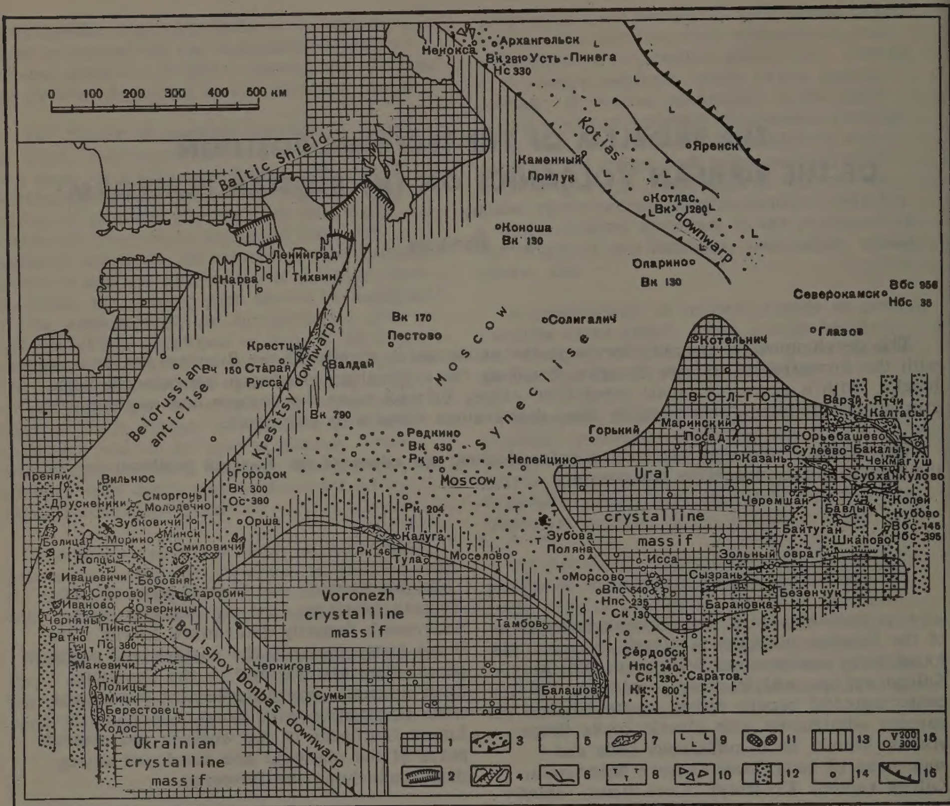

FIGURE 1. Distribution of volcanics throughout the Russian platform, at the end of Riphean time.

1 -- differentially uplifted segments of crystalline basement, with the Riphean either not deposited or very thin; 2 -- grabens with Jotnian formations; 3 -- troughlike downwarps with the most complete Riphean sections; 4 -- horsts with Riphean deposits either lacking or very thin; and grabens with the most complete Riphean sections (southwestern Belorussia, western Bashkiriya); 5 -- syneclises and anticlises with relatively incomplete Riphean section; 6 -- gabbro-diorites and gabbro-norites veins, dikes and stocks; 7 -- bedded bodies of andesite-basalts, basalts, diabases, spilites; 8 -- andesite-basalt, basalt and diabase tuffs and tuffites; 9 -- ash tuffs; 10 -- tuff breccias of alkaline basalts; 11 -- magnetic and gravity maxima, interpreted as gabbro and basalt bodies.

Zones of alleged volcanic formations: 12 -- in the structures of southwestern Belorussia and western Bashkiriya; 13 -- on flanks of troughlike downwarps; 14 -- boreholes; 15 -- thickness of Riphean deposits (in meters); 16 -- boundary of the Timan' downwarp.

Symbols: Вк -- Valday sequence; Ос-- Orsha formation; Нс -- Nenoksa formation; Впс -- upper Pachelma series; Нпс -- lower Pachelma series; Ск -- Serdobsk sequence; Кк -- Kaverino sequence; пс -- Pinsk formation; Вбс -- upper Bavly formation; Нбс -- lower Bavly formation; Рк - Redkino sequence.

to the Paleozoic group. N.S. Shatskiy [30] believes the Redkino and Valday deposits to be Proterozoic-Riphean, whereas B.M. Keller [8] assigns to the latter only the Redkino sequence, relegating the overlying formations to the Lower Cambrian.

Three types of the Riphean section are recognized on the Russian platform.

The first type includes stratigraphically more-complete sections with the Kaverino and Serdobsk sequences and lower Pachelma series. They are developed in the Kaverino and Serdobsk areas of the southeast part of the Pachelma downwarp. Here also belong

sections including the Serdobsk and lower and upper Pachelma rocks of the northwest part of the Pachelma downwarp, and the Valday and Orsha rocks of the Kresttsy downwarp; the Redkino and Valday rocks north of the Voronezh massif; the Nenoksa and Valday rocks in the Kotlas downwarp; both of the Bavly formations of western Bashkiriya; and the Pinsk and Valday rocks in the southwestern Belorussian depressions.

The second type comprises sections made up of a thin sequence of Valday rocks, within the Moscow syneclise and Belorussian anticlise.

Finally, the third type is characterized by the lack of Riphean deposits.

Massive Riphean deposits are either totally lacking on the Baltic shield, Voronezh-Ukrainian, and Volga-Urals crystalline massifs, or else are very thin. This is true also for the Riphean on basement highs of the southwestern Belorussia and western Bashkiriya. Table 1 gives the best examples of the Riphean-type sections.

TECTONIC STRUCTURE OF THE RIPHEAN VOLCANICS ON THE RUSSIAN PLATFORM

Recent work [1, 3, 25-31] has established that the close of the Riphean on the Russian platform witnessed the following principal tectonic elements.

A. Differentially uplifted segments of the crystalline basement: the Baltic shield, and the Voronezh, Ukrainian, and Volga-Urals crystalline massifs.

B. Onega and Ladoga grabens filled by Jotnians formations.

C. Grabenlike downwarps: the Pachelma, Kresttsy, Kotlas, and west Ukrainian. Full and diversified Riphean sequences rest upon the deeply buried crystalline basement of these downwarps.

D. The southwestern Belorussian and western Bashkirian tectonic forms consisting of basement highs with the Riphean either not deposited or very thin, and of depressions containing the most complete sections.

E. The Moscow syneclise with the Valday rocks resting upon the crystalline basement.

F. The Belorussian syneclise with a comparatively small thickness of the Valday.

The distribution of these structural elements is illustrated on the index map (Fig. 1) compiled from a pre-Devonian paleogeographic map [27] and complemented by deep drilling and geophysical data. The outlines of basement highs are mostly approximate, since they are covered by a sedimentary mantle.

Grabenlike, or more correctly troughlike, downwarps such as the Kresttsy, Pachelma, and the supposedly contemporaneous Bol'shoy Donbas downwarp, have split the platform basement into smaller blocks. The Sarmatian pre-Riphean shield was split into the Volga-Ural, Voronezh, and Ukrainian crystalline massifs. A similar differentiation process appears to have taken place within the Baltic shield where the depressions are filled with sparagmite deposits. At the beginning of the Riphean, the present Moscow syneclise also presented a shield where the narrow Kresttsy downwarp originated and developed.

THE KRESTTSY TROUGHLIKE DOWNWARP

Wells drilled near the village of Kresttsy, in the downwarp of that name, penetrated ferrous Proterozoic sandstones and Archean biotite-hornblende gneisses, at depths of 1,830 and 1,743 m below sea level. At Staraya Russa, to the west, the Archean lies at 900 m below sea level, whereas analogous formations in the Valday to the east were found at 1,500 m below sea level.

A.N. Geysler [3] has shown that the Archean gneisses and Proterozoic quartzites of the Kresttsy downwarp are overlain by coarse-grained tuffaceous sandstones alternating with tuffites whose clastic material consists of magmatic basic rocks, and is represented by diabase, diabase porphyry and vitreous pumicelike rocks, lapilli, and ash. This sequence carries diabase seams from several centimeters to 0.75 meter thick. The overall thickness of the volcanic-sedimentary rocks is 430 to 490 m. They are followed by sandstones and, still higher up, by laminarites-carrying shales. The total thickness of this ancient sedimentary complex at Kresttsy is 760 to 790 m. These figures suggest that sedimentation was occurring here under conditions of more intensive downwarping than in the adjacent areas. At Staraya Russa, the Valday beds are 150 m thick, and 170 m at Pestov [12]. The Kresttsy downwarp is narrow, not more than several tens of kilometers wide, and apparently extending for many hundreds of kilometers. It is asymmetric with a steeper western slope.

The southwest extension of the Kresttsy downwarp is a depression filled with Orsha (O) and Valday (V) deposits (Orsha, Gorodok, Smilovichi, Minsk). Here, the Archean crystallines are directly overlain by red sandstones,

nearly pure quartz in most instances, devoid of argillaceous contamination and interbeddings, with quartz cement, locally very tough. Toward the top, sandstones are replaced by argillaceous silts, as much as 50 m thick, carrying dolomite beds (1 to 2 m) of a peculiar plicated texture. Thickness of the Orsha formation ranges from 75 to 380 m [1].

Tuffaceous deposits in the lower Valday section were penetrated by deep drilling in the area of Smilovichi and Minsk. Valday rocks lie above the 264-m thick Orsha formation, at Smilovichi. The section here begins with coarse terrigenous rocks, with quartz-feldspar sandstones and siltstones above them, carrying various amounts of volcanic material. The clastic component of these deposits consists essentially of extrusives including volcanic glass. Volcanic material is present also in the cement where it is highly chloritized. Volcanic deposits are 66 m thick at Smilovichi, and 33 m thick at Mins. The overall thickness of the Gdovsk beds at Smilovichi is 232 m. Volcanic formations gradually change upward to the nonvolcanic capped by dark-gray siltstones (89 m) correlative with the laminarite. The total thickness of the Orsha, Gdovsk, and laminarite beds, at Smilovichi, is 585 m [13, 14].

An asymmetric downwarp appears to have existed, at the beginning of the Riphean, between the Voronezh high and the Moscow pre-Riphean shield. Unfortunately, data on its boundaries are inadequate, as yet. However, as a result of detailed study of ancient sections (from well cores and cuttings) in the areas of Redkino, Povarovo, Moscow, Serpukhov, and Kaluga, A.V. Kopeliovich concluded that "deep submersion occurred along the site of the present northeast slope of the Voronezh massif, during a time corresponding at least to that of the Redkino deposition" [10, p. 58].

It is not impossible that this area of deep submergence is related to the Pachelma downwarp, in the southeast, and opens into the Kresttsy downwarp, to the west. The first rocks described by A.V. Kopeliovich north of the Voronezh massif [9] are those of the Redkino area. The study of this section made it possible to gain a better understanding of the age relationship between the barren terrigenous sequences of the Baltic and Moscow areas, and thereby making it a standard section.

Crystalline rocks and the ancient weathered crust, at Redkino, are overlain by the following formations:

1. Sandstones, light gray, coarse grained, with argillaceous and dolomitic cement; thickness, 12 m.

2. Shales, dark gray with glauconitic and dolomitic layers. Shales and siltstones of the lower part have a peculiar texture reminiscent of submarine creeps; thickness, 84 m. A.V. Kopeliovich [9] gave the name, Redkino sequence to the identifiable sandy and shaly rocks of this sequence.

3. Higher in the section, there lies a primarily sandy sequence, 86 m thick.

4. Thin-bedded siltstones and ferrous shales, chocolate brown to gray green, thin to extremely thin bedded. Siltstone cement is mica-argillaceous, frequently dolomitic. There are subordinated intercalations of fine to mixed-grained polymictic sandstones with argillaceous, locally dolomitic, cement. Their clastic material is poorly sorted and consists of quartz, feldspar, fragments of assorted extrusives, and quartz-feldspathic, and thin mica-argillaceous vein rocks. Kaolinite, dolomite, and barite are common in the cement. Films of Laminarites antiquissiums Eichw. occur along bedding planes. Spores, Leiotriletes hialinus Naum., Trachytriletes laminaritus Naum., T. solidus Naum., have been observed in the lower part of this sequence. Its total thickness is 345 m.

Beds 3 and 4 are correlative with the Valday sequence of the northwest areas of the Russian platform.

5. Sandstones, gray to greenish gray, 13 m thick.

6. Shales, greenish gray, thin bedded, 42 m thick, carrying remains of tubular worms, Sabellidites cambriensis Jan.

Areno-argillaceous deposits of beds 5 and 6 correspond to the Baltic sequence. All students agree on its Early Cambrian age. The overall thickness of the Redkino and Valday sequences is 527 m.

A.V. Kopeliovich [10] has shown that the body of argillites, comprising the bulk of the Redkino sequence is clearly traceable in the Redkino, Povarovo, Moscow, and Serpukhov boreholes. It thickens gradually from north to south: it reaches 84 m at Redkino, 128 m at Povarovo, 182 m in the Moscow section, and 204 m at Serpukhov. Its thickness on the Voronezh massif is zero. Thus, the maximum downwarp of Redkino time occurred approximately at the latitude of Serpukhov. North of there, the Redkino sequence thins appreciably. Thinning of the total pre-Devonian section toward the Voronezh massif, on the other hand, is not because of its wedging out, but rather progressive truncation by younger formations.

An outstanding feature of this area is the presence of tuffaceous deposits at the junction

of the Voronezh crystalline massif and the downwarp to the north. According to S.V. Tikhomirov [20], the Archean granite-gneisses in the Mstikhino area are overlain by a coarse breccia with fragments of granite gneiss and other metamorphic and extrusive rocks, with an arkosic cement. Their thickness is 14 m. They are followed by black tuffs (1.5 to 2 m) approaching alkaline andesitic lavas in composition, and carrying numerous grains of albite and quartz. The tuffs are overlain by gray porous tuffites (86 m), with granite-gneiss fragments in their lower portion. The composition of rocks above the tuffites (15 m) is unknown. Above them lie Devonian rocks.

THE PACHELMA DOWNWARP

The dating of this downwarp depends on the age determination of its fill. We are still in the dark as to the stratigraphic relationship of this section with the Jotnian formations of the Onega region, and as to the lower boundary of the Riphean. N.S. Shatskiy [31] has shown that, toward the close of the Riphean, the Pachelma downwarp was fringed, both to the east and to the west, by crystalline massifs of the Sarmation shield.

The sequence above the crystalline basement of the most depressed southeast part of the Pachelma downwarp is as follows:

1. Red sandstones, coarse-grained quartz, with beds of gravel and conglomerate in the upper part; their thickness at Kaverino is more than 800 m. These deposits belong to the Kaverino sequence [17]. The Kaverino rocks are more than 400 m thick at Serdobsk.

2. Sandstones, greenish brown, glauconitic; quartz-feldspathic, about 50 m thick.

3. Dolomites, light gray, microcrystalline, very dense, 80 m thick.

4. Red dolomitic marls, fine-grained sandstones and drab-red shales; total thickness, 100 m. I. Ye. Postnikova [17] draws the upper boundary of the Serdobsk sequence on the top of this bed.

5. Red polymictic sandstones, and black shales, very tough, about 75 m thick.

6. Alteration of thin, greenish gray quartz-feldspathic sandstones, locally glauconitic, and dark gray marls, siltstones, and black shales. The cement in sandstones and siltstones is glauconitic and dolomitic. The overall thickness of these rocks reaches 150 m.

N.S. Shatskiy [31] combines red bed 5 and dark-colored bed 6 into the lower Pachelma series. They are conconformably overlain by sandstones with a weathering crust, followed by Middle Devonian formations. The overall thickness of this Riphean section exceeds 1,200 m. It is shown in Table 1, which also shows some 900 m of an ancient sedimentary section of the Pachelma area.

It is notable that the Kaverino correlatives (bed 1) are missing in the Pachelma area. On the other hand, beds overlying the lower Pachelma, which are nearly missing in the Serdobsk section, are much thicker here. Thus, the upper Riphean interval is 775 m thick at Pachelma, and 225 m at Serdobsk (Table 1). This lack of uniformity in the downwarping magnitude of different segments at different periods of the Riphean is a characteristic feature of not only the Pachelma but of other grabenlike downwarps, especially the Kresttsy. Thick Orsha (O) deposits are developed in the southwestern part of the latter, and the Valday (V) developed in its northern and northeastern part.

It is of importance that Riphean tuffs and tuffaceous sedimentary formations are formed in the section comprising the northwest part of the Pachelma downwarp (at Mosolovo, Morsovo, Zubova Polyana). According to Z.P. Ivanova, most of the tuffaceous and extrusive material is observed in the Mosolovo borehole section where it occurs at 1,425 to 1,575 m. This section is located on the slope between the Voronezh massif and Pachelma downwarp. Its lower part consists mostly of sandy deposits; the upper, mostly of shales [22]. Given below is a generalized description of this section, reading upward.

1. Sandstones, mixed and coarse grained, locally with gravel; quartz to quartz-feldspathic with much micas and ore minerals; and an opal, chlorite, silica, and ferrous cement. This sequence contains beds of siltstones and tuffaceous sandstone, 6 m thick, consisting of angular fragments of volcanic glass and strongly disintegrated extrusives cemented with chlorite. The tuffaceous siltstones also contain strongly disintegrated extrusive fragments. Total thickness, 140 m [1].

2. Shales with some intercalations of tuffaceous siltstones and sandstones. Argillaceous rocks are reddish brown to greenish gray, micaceous, microstratified in some beds, with thin lenses of siltstones consisting of quartz, feldspars, and mica, cemented with a chloritic or silica argillaceous substance. Sandstones are medium to fine grained, well sorted, quartz to quartz-feldspathic. Tuffaceous varieties are 70 to 80 percent isotropic grains of volcanic glass and disintegrated extrusive fragments cemented with a chloritic substance. Total thickness, 2 to 60 m.

Table 1

RIPHEAN SECTION TYPES

(After A. V. Kopeliovich [9], Ye. M. Lyutkevich and M. I. Peysik [12], I. Ye. Postnikova [18] and N. S. Shatskiy [31])

Relatively complete sections									Incomplete sections			Riphean deposits are missing
Serdobsk			Pachelma			Redkino			Porkhov			Shields and basement highs in the structures of southwestern Bellorussia and Bashkiriya
Devonian sediments												
Upper Pachelma	Sandstones	Thickness 30 m	Upper Pachelma	Alternated thin black shales and gray sandstones and siltstones	Thickness 240 m	Valday sequence	Alternation of thin chocolate-brown shales and gray-green sandstones and siltstones. Films of Laminarites antiquissimus Eichw.	Thickness 430 m	Valday sequence	Sandy shales with Laminarites antiquissimus Eichw.	Thickness 150 m	—
				Multicolored sandstones, quartz-feldspathic	Thickness 300 m							
Lower Pachelma	Alternation of thin sandstones, marls, siltstones, shales; dark gray, glauconitic. Red	Thickness 225 m	Lower Pachelma	Alternated thin gray-green sandstones, siltstones, shales. Red sandstones	Thickness 235 m		Glauconitic sandstones.			Sandstones and sands		
Serdobsk sequence	Marls, sandstones, red shales; Dolomites: Glauconitic sandstones	Thickness 230 m	Serdobsk sequence	Dolomite, shales Glauconitic sandstones	Thickness 130 m	Redkino sequence	Shales with dolomite intercalations Sandstones	Thickness 95 m		—		—
Kaverino sequence	Sandstones, gravels, conglomerates	Over 400 m	Kaverino sequence	Missing			—			—		—
Archean formations												

3. Tuffs and tuffites made up of fragments of extrusives, chlorite, chloritized micas, with a small amount of quartz and feldspar, all cemented with a silica-chloritic to dolomitic substance. Porphyritic texture is observed in some volcanic varieties, with feldspathic incrustations. Such a texture suggests the presence of extrusives subjected to intensive secondary alterations. Total thickness of this bed is 3 to 21 m.

4. Alternation of thin fine-grained rocks -- slaty shales, shales and siltstones. All colored brown-red to brown, with a layer of blue-gray slaty shales (20 m), in the middle part, alternating with thin-bedded, humus-sapropelic slates. The same layer includes two thin dolomitic-sideritic layers. Slaty shales of the upper part exhibit plicated structure. Thickness, 4 to 194 meters. Z.P. Ivanova notes the films of organic matter in these ancient sediments. This is especially true for the upper part of this section.

As correlated from the Mosolovo, Morsovo, Zubova Polyana, and Nepeytsino boreholes, the Riphean interval shows a decrease in the amount of volcanic material, going away from the Voronezh massif. Thus, the Morsovo section contains a substantially thinner volcanic sequence. It is still thinner in the Nepeytsino and Zubova Polyana well sections, located on the western slope of the Volga-Ural basement high. In the Gor'kiy area, volcanic glass is found only in the light fraction of clastic rocks penetrated at depths of 1,595 to 1,600 m.

The above data on the types of sections filling the Kresttsy and Pachelma downwarps show that these represent a single system of narrow downwarps separating large blocks of the crystalline basement.

KOTLAS DOWNWARP

Along with depressions in the body of shields, the close of the Riphean witnessed the formation of downwarps along the edge of the pre-Riphean shields, which also may be considered among the grabenlike troughs.

The Kotlas downwarp fringes the northeast rim of the Moscow pre-Riphean shield, with a northwesterly trend. Ite existence is inferred from the following data. A sequence of multicolored quartz to quartz-feldspathic sandstones with a ferrous, ferrokaolinite and quartz cement rest upon the crystalline basement, in the Nenoksa area, which is the northwest part of the downwarp. This sequence is 333 m thick, It is classified by A.I. Zoricheva [6] as the Nenoksa formation. Its eroded surface is overlain by basal conglomerates and sandstones with beds of gravels and green shales (27.5 m).

These coarse-grained sandstones are followed by greenish-gray shales interbedded with lighter colored mica-quartz siltstones. The students of these rocks note the presence of a thin horizontal stratification, with the bedding planes generally covered with a dark organic film (laminarites). An important feature of these laminarite beds in the Nenoksa section and elsewhere in the Kotlas downwarp is the presence of thin intercalations of montmorillonite clays. The latter are regarded as ash deposits altered by sea water. The laminarite beds are 254 m thick, with their upper boundary drawn conditionally on the appearance of multicolored sandy rocks which are assigned to the Baltic complex. The overall thickness of this ancient sedimentary sequence at Nenoksa is 615 m.

The crystalline basement plunges to the southeast, and the Riphean section thickens in the same direction. In the Ust'-Pinega area, the total thickness of ancient sediments attains 800 m; still farther southeast, an incomplete Valday section is more than 1,200 m thick (Kotlas).

A borehole in the Kotlas area penetrated 2,300 m of sediments without reaching the crystalline basement. According to A.I. Zoricheva [6], the lower part of this section is represented by more than 500 m of blue-green siltstones and sandstones. They are overlain by thin-bedded laminarite shales with a millimeter-thick montmorillonite laminae and intercalations of siltstones carrying microfragments of extrusives which are partially to fully replaced by montmorillonites. The laminarite beds there are 786 m thick.

These sections characterize the structure of the southeast part of the Kotlas downwarp, as shown in Fig. 1. Riphean deposits grow thinner to the southeast: the Valday sequence is 130 m near Oparino.

The northeastern Kotlas downwarp boundary has not been determined as yet. Ancient rocks of that area were penetrated by a single borehole, near Yarensk, and that not fully. Laminarite beds in the Yarensk borehole are 235 m thick, as compared to 786 m in the Kotlas area. The reason for this shortening is not clear. It is possible that the Kotlas downwarp is a very narrow trough, with the Yarensk section located on its eastern slope. It also is not impossible that this downwarp is built of differentially uplifted basement blocks, with Riphean sections of various thicknesses. Other possibilities should not be ruled out.

WESTERN UKRAINIAN DOWNWARP

This feature, located along the west rim of the Ukrainian crystalline massif, appears to belong to the grabenlike group. Riphean deposits here, according to O.V. Krasheninnikova [11] exceed 1,000 m in thickness, being represented by sedimentary and volcanic rocks of the Tashkovo, Gorbashev, Izyaslav, and Ushitsk formations [11].

Volcanics of the Goryn' basin make up a large portion of the Izyaslav formation, Ostrog series. They are penetrated by a number of boreholes located in a narrow belt trending northwest between the villages of Izyaslav, to the southeast and Politsy to the southwest. Within this belt, extrusive-sedimentary rocks rest upon the Gorbashev formation comprised of multicolored arkosic sandstones interbedded with siltstones and shales. The penetrated thickness of the Gorbashev formation is 40 to 50 m. It is followed by the Babino beds making up the lower part of the Izyaslav formation. These beds are represented by sandy and pebbly tuffites interbedded with volcanic tuffs, tuffaceous sandstones and siltstones. Red colors predominate in the lower part of the section; gray-green, in the upper. The middle part contains an 8 to 10 m-thick bed of ill-sorted sandstones abundant in coarse clastic material, chiefly of extrusive pebbles.

Tuffaceous sediments of the Babino beds are characterized by a peculiar composition: they consist to a considerable extent of analcite, serpophite, and chlorite. Among accessory minerals, there is ilmenite, magnetite, titano-magnetite, and zircon and apatite grains. Analcite with some hydrous silicates form the cement. They also fill the pores and replace grains of clastic material. Intensively metamorphosed rocks are marked by their high content of Mg and Ti oxides, and by the presence of such elements as Ni, Co, Cr, Cu, Pb, I, and Sc.

The Babino beds are overlain by the Sergiyevsk beds which make up the upper part of the Izyaslav formation. The Sergiyevsk beds begin with tuffaceous siltstones and sandstones, from 4 to 11 m thick. Paleobasalts, penetrated by boreholes at the villages of Khodosya, Zhil'zhe, and Berestovets rest upon an eroded surface of these beds, or directly on the Babino beds, as along the lower course of the Gorn' River and elsewhere. The thickness of these basalts ranges from 10 to 25 m. Conglomeratic sandstones and siltstones are developed at their top. In the middle and upper courses of the Goryn', the sub-basalt rocks are overlain by multicolored tuffaceous shales, siltstones, and sandstones, with dark gray and dark brown argillite mudstones with pyrite lentils. The thickness of these rocks ranges from 10 to 25 m. The upper part of the Sergiyevsk beds is represented chiefly by shales and argillite mudstones with lenticular intercalations of light gray calcareous sandstones, siltstones, and sandstones. Individual layers are inconsistent in thickness, locally wedging out completely.

In the Goryn' basin, the Izyaslav formation is followed by the Ushitsk formation of mudstones, shales, siltstones, and sandstones.

These ancient deposits are distributed along the west slope of the Ukrainian crystalline massif. Going west, they disappear underneath younger formations.

TECTONIC FORMS OF SOUTHWESTERN BELORUSSIA

Toward the close of the Riphean, southwestern Belorussia presented a series of relatively small basement highs separated by depressions filled with ancient sediments (Fig. 1). These deposits contain a consider-amount of volcanic material.

In southern Belorussia, the crystalline basement rocks are overlain by red, fine-grained, quartz-feldspar sandstones and coarse-grained siltstones of the Pinsk formation. These rocks are missing on the Mikashevichi and Bobovnya basement highs. Within the Ratno high, the Pinsk formation is 40 m thick; it is 380 m thick in an adjacent depression to the east. On the west slope of this depression, the Pinsk formation is overlain by sandstones, shales, and diabases, correlated by A.S. Makhnach [13] with the lower part of the Valday sequence north of Belorussia.

West of theRatno high, boreholes penetrated 116 m of basic extrusives without reaching their base. The rocks are chiefly dark gray to black paleobasalts with a greenish cast and are dense, massive, usually fully crystalline, locally aphanitic. Some varieties have an amygdaloid structure. Subordinate rocks are represented by spilites, tuffs, and tuff breccias. Higher up, there are tuffaceous-sedimentary, areno-argillaceous rocks, about 45 m thick, consisting of poorly sorted sandstones, arkosic, in many places gravelly, tuffaceous with a mica-argillaceous, ferrous and volcanic, locally chloritic and carbonate, cement.

Table 2 shows the relationship of these rocks.

Wells drilled in the area of Ozernitsy, Gavril'chitsy, Starobino, and Glusska (Luchki) reveal the composition of rocks in the

Table 2

Riphean Sections Resting on the Ratno-Basement High and in the Adjacent Depressions, to the East and West. (After A.S. Maknach [13])

Downwarp west of the Ratno high	Ratno-basement high	Downwarp east of the Ratno high	
Chernyany	Ratno	Ivanovo	Pinsk
Cretaceous deposits			
Valday sequence: Tuffaceous sandstones, siltstones, shales, breccias, tuff breccias, tuffs, tuffites, paleobasalts, spilites; Over 160 m	Valday sequence: Missing	Valday sequence: Diabases, sandstones, shales; Thickness 30 m	Valday sequence: Missing
	Pinsk formation: Red sandstones; Thickness 40 m	Pinsk formation: Red sandstones, siltstones, shales; Thickness 100 m	Pinsk formation: Sandstones, shales, siltstones, quartz-feldspathic, pink, red-brown; Thickness 380 m
	Archean: Biotite-plagioclase gneisses		Archean: Granodiorites

lownwarp between the Bobovnya and Mikhahevichi crystalline-basement highs. In the itarobino area, tuffaceous-sedimentary rocks ıre penetrated below the Devonian. They are uffaceous sandstones, 100 m thick, cemented vith chlorite and carrying coarse fragments of extrusives including volcanic glass. They ıre underlain by sands and poorly cemented ırkosic, pink to brown-red sandstones and iltstones (P), 330 m thick. Granodiorites ıre penetrated at the base of the section. Riphean deposits are missing on the Mikahevichi- and Bobovnya-basement highs.

Data on the distribution of the various Riphean sections throughout southwestern Belorussia are still scant. However, they point to the fact that tectonic forms of this region, toward the close of the Riphean, presented several small uplifts separated by depressions filled with deposits of the Pinsk formation and with volcano-sedimentary rocks correlative with the Valday sequence.

TECTONIC FORMS OF WESTERN BASHKIRIYA

A study of the relationship between the Volga-Ural crystalline massif and Riphean deposits developed along its periphery reveals peculiar tectonic features along its southeast slope, similar to western Belorussian structures. The southeastern edge of the Volga-Ural massif is characterized by its uneven outline, with horstlike basement highs separated by deep depressions, as shown on Figure 1. To the west, they merge into the single Volga-Ural massif, whereas in the east, north, and south they are separated from each other by deep depressions with comparatively complete Riphean sections. For instance, the lower Bavly formation is best developed in a depression located in the Bavly area [20], where the lower multicolored sequence, as much as 28 m thick, rests upon the crystalline basement. It is followed by a red dolomitic-terrigenous sequence (82 m) and

then by the middle red sandstone sequence, as much as 285 m thick. The total thickness of the lower Bavly formation (Lb) is 395 m.

The upper Bavly formation (Ub) is incompletely represented in this structure. Its upper part ("upper interbedded sequence") is missing in some of the sections, with the underlying upper sandstone sequence also missing in some other sections where the total formation is but 145 m thick.

The overall thickness of Riphean deposits in this depression is 540 m. They are missing on uplifts to the north, southeast, and west.

Lower Bavly rocks filling deep grabenlike depressions carry volcanics. The latter are best represented at the village of Or'yebasheva, in the Kaltasy area where, according to A.M. Dymkin, L.F. Solontsov, and S.S. Ellern [4], two volcanic layers are observed among the sediments. The lower layer is associated with the terrigenous-dolomitic sequence and is fully penetrated at a depth of 2,313 to 2,388 m. The upper layer is 20 m thick. It is underlain by the terrigenous-dolomitic sequence and is overlain by deposits relegated to the "lower interbedded sequence" (Ub). The lower volcanic layer is represented by standard diabases characterized by plagioclases of the labradorite type, monoclinic pyroxenes, and possessing an ophitic texture.

Extrusives of the diabase and diabase porphyrite types are identified in many other localities within the Bashkirian downwarp. At the Varzi-Yatchi settlement, they are represented by amygdaloid porphyrites below gray shales of the "lower interbedded sequence." At Staro-Petrovo near Birsk, a diabase layer (3 m) divides the terrigenous-dolomitic sequence, which is as much as 205 m thick. The diabase consists chiefly of a labradorite type plagioclase, augite, and ore minerals, most commonly magnetite. Its texture is ophitic. Rocks of the same composition are encountered in Baikbashevo.

A comparison of the southwestern Belorussian with western Bashkirian structures reveals that they are characterized by the same types of ancient deposits. In both, the Riphean is best developed in depressions and, as a rule, is either missing or very much abbreviated on highs. A common feature of both is the presence of paleobasalts, diabases, and spilites.

MOSCOW SYNECLISE

The broad and flat Moscow syneclise extends between the Baltic shield to the north and the Volga-Ural high to the south. It borders on the narrow Kresttsy downwarp to the west and southwest, and on the Kotlas downwarp to the northeast. Table 3 shows the comparative thicknesses of Riphean deposits developed throughout the Moscow syneclise and the adjacent downwarps.

The Archean basement of the Moscow syneclise is overlain by Riphean deposits penetrated by peripheral boreholes in the area of the villages of Pestovo, Konosha, Oparino, and Gor'kiy. They were penetrated in part, at Shar'ye. The nature of the Riphean deposits is unknown in the central part of the syneclise where the boreholes stopped in younger formations. Along its periphery, Archean crystallines are overlain by sandstones, from several to 30 m thick. The bulk of the section is made up of shales, as much as a few hundreds of meters thick: about 90 m at Gor'kiy, 170 m at Pestovo, 130 m at Konosha, and 130 m at Oparino.

Redkino rocks have not been observed in the above-named sections. They are either missing or so thin as to be undistinguishable among the lower Valday rocks. It appears that, prior to the Valday deposition, the site of the Moscow syneclise was a high similar to the Sarmatian and Baltic shields.

BELORUSSIAN ANTICLISE

This feature was located, at the close of the Riphean, between the Baltic shield to the west and the Kresttsy downwarp, to the east. A comparatively small thickness of Valday deposits was laid down upon its surface. Neither the Orsha-type deposits nor the Redkino sequence are recognized here. According to Ye. M. Lyutkevich and M.I. Peysik [12], in the Porkhovo area the overall thickness of Gdovsk and laminarite beds is 151 m; 190 m at Staraya Russa; 273 m at Nevel'; and 105 m at Vil'nyus.

The ancient section in these areas is subdivided into two parts: chiefly arenaceous Gdovsk beds below, and chiefly shales, more or less silty, with subordinate siltstones and sandstones (laminarite beds), above. Their bedding places usually exhibit sapropelite films of the Laminarites antiquissimus Eichw. type.

* * *

Given above are Riphean type sections and an outline of their distribution, characterizing various structural elements on the Russian platform. The onset of the Riphean witnessed the formation of narrow, troughlike downwarps

Table 3

Riphean Sections Within the Moscow Syneclise and the Kresttsy and Kotlas Downwarps. (After A. N. Geysler [3], A. I. Zoricheva [6], Ye. M. Lyutkevich, and M. I. Peyski [12])

Kresttsy downwarp			Moscow syneclise			Kotlas downwarp		
Kresttsy			Pestovo			Kotlas		
Cambrian deposits								
Valday sequence	Shales, sandstones. Tuffaceous sandstones. Tuffites, tuffs, diabases	Thickness 790 m	Valday sequence	Sandy shales	Thickness 170 m	Valday sequence	Thickness more than 1285 m	Laminarites, shales with montmorillonite intercalations. Sandstones, siltstones
Archean	Biotite-hornblende gneisses		Archean	Granites				

in the platform basement, which were then filled with terrigenous and terrigenous-dolomitic sediments. The Pachelma downwarp received the Kaverino and Serdobsk deposits; Redkino beds were laid down north of the Voronezh massif; Nenoksa sandstones accumulated in the Kotlas downwarp, with the Orsha formation in the southwest part of the Kresttsy downwarp. The Pachelma and probably the Bol'shoy Donbas downwarps split the Sarmatian shield into the Volga-Ural, Voronezh, and Ukrainian crystalline massifs. At the same time, the junction of the Bol'shoy Donbas, Kresttsy, and western Ukrainian downwarps, in southwestern Belorussia, marked the subdivision of the crystalline basement into several small highs and lows, the latter filled with deposits of the Pinsk formation. A similar structural setup originated in western Bashkiriya, where the depressions were filled with the lower Bavly formation.

During the Valday deposition, the Moscow pre-Riphean shield became a shallow syneclise, while considerably thicker sediments continued to accumulate in the Kresttsy, Pachelma, and Kotlas grabenlike downwarps. Valday deposits were either not laid down at all on the Ukrainian, Voronezh, and Volga-Ural crystalline massifs, or else were very thin.

These structural features of the platform, during the Riphean, determined the condition of origin and the initial tectonic stage of development of its sedimentary mantle.

PATTERNS IN THE SPATIAL DISTRIBUTION OF THE VOLCANIC ROCKS

Two groups of magmatic formations are recognized among the Riphean deposits: tuffaceous and extrusive. Stratified bodies of dolerite, diabase, diabase-porphyrite and spilite lavas are concentrated along the slopes of the grabenlike downwarps as well as in structures of southeastern Belorussia and western Bashkiriya. Tuffaceous deposits are much more widely distributed than the lavas. In some places, tuffs and tuffites, too, are distributed along the flanks of the depressions, as is the case along the west slope of the Kresttsy downwarp. Elsewhere they fill the entire depression, as in the northwest part of the Pachelma and Kotlas downwarps (Fig. 1).

The previously mentioned lavas at Or'yebashevo, Staro-Petrovo, Baikbashevo, and Yarzi-Yatchi are confined to slopes of the western Bashkirian grabenlike depressions. Similar formations are present at the junction of the Voronezh crystalline massif and adjacent downwarps. Here belong andesite-basalts near Kaluga and the Balashev basalts.

In southwestern Belorussia, basalt rocks are developed along the peripheries of basement highs, occupying a definite tectonic position along slopes of differentially uplifted and depressed basement segments. Basic lavas were penetrated in drilling along the east rim of the Ratno uplift: at Ivatsevichi, Sporovo, Khomsk, Ivanovo, and Manevichi; also along its west rim, in a number of boreholes in the Malority and Divino areas (Table 2; Fig. 1). In addition, there are indirect data on the distribution of volcanic rocks in this region. There are several known magnetic and gravity anomalies in the meridional belt between Grodno and Brest, believed to be caused by deep-seated basaltic volcanics. The latter, as previously noted, were penetrated by wells in the Malority and

Divino areas [26].

Riphean volcanic activity also occurred in the southwest of the Russian platform, along the west rim of the Ukrainian crystalline massif. This is suggested by extrusive formations of a paleobasalt type, among Riphean deposits in the area of Mitsk, Politsy, Berestovets, Khodos, and elsewhere.

Besides other extrusives, Riphean volcanics appear to include vein rocks, represented by gabbro-diabases and gabbro-norites. The data on hand suggest that gabbro-diabase vein bodies are associated with the west edge of the Volga-Ural basement high, more specifically with the zone of transition to structures of the western Bashkirian horst types. Here, the east edge of the crystalline basement, too, is split into horsts and grabens filled with deposits of the upper and lower Bavly formation (Fig. 1). Gabbro-diabases have been found in many localities along the east edge of the Volga-Ural massif, where they lie below Devonian sediments in the area of the villages of Suleyevo, Bakalovo, Chekmagaush, Aznakayevo, Subkhankulovo, Bishindy, and other places. In the village of Kopey-Kubov, they were penetrated below the Bavly formation. Gabbro-norites are developed in the same zone and were penetrated by drilling in the Cheremshan and Samarskaya Luka (Samara Bend) area: at Syzran', Zol'nyy Ovrag, Pechersk, and elsewhere.

Coincident with these gabbro-norite zones, there are amphibolites which appear to be, in many instances, alterations of gabbroid rocks; amphibolites from Baranovka, Zol'nyy Ovrag, and Baytugan. These areas of basic volcanics are considerably scarcer in the interior of the Volga-Ural massif. They are known there only at Mariinskiy Posad and north of Kazan'.

As already noted, deep-seated basic rocks are also fairly well developed in southwestern Belorussia, where they are found in the north part of the Bobovnya-basement high (villages of Zubkovichi, Morino, Koptsy, and Belitsa), on the Mikashevichi high and in a number of other places, such as Pishe, Druskeniki, Prenyay, Smorgon', Molodechno, etc. The very number of these localities indicates a concentration there of basic rocks. The manner of their occurrence is not as yet clear. The Zubkovichi-Morino gabbro body is identified with a marked magnetic anomaly, of the order of 3,000 gammas. It was believed that this anomaly had been induced by deep-seated iron ores, until wells drilled in Zubkovichi and Morino penetrated a crystalline basement of strongly limonitized pyroxene-biotite gabbros.

This occurrence provides a reason to believe that other magnetic and gravity maxima in the narrow, northeast-trending belt between Slutsk and Orsha also reflect deep-seated gabbros [26].

There are not enough data on the manner of occurrence of these gabbros. What data there are, suggest that gabbro bodies do not extend over large areas but occur rather in a network of veins, dikes, and stocks similar to those of the Siberian platform [15, 16]. It is quite probable that a temporal as well as spatial relationship exists between vein occurrences of gabbro-diabases and stratified lava bodies. Gabbro-diabases which occur among granite-gneisses utterly lack any gneissoid characteristics although they rest at the same level with the enclosing metamorphics [2]. This indicates a considerably younger age for the gabbro-diabases. At Or'yebashevo, gabbroid rocks beneath Devonian sediments (boreholes 14, 23) are found near diabases and spilites developed in different layers of the Bavly formation (borehole 57). The same rock relationship is observed at Staro-Petrovo where gabbro-diabases are penetrated at the base of the Bavly formation, with doleritelike rocks penetrated 130 m higher in the section. All of this, leads to a belief that a single magma reservoir was the source of both the extrusives and intrusives, and that the different forms of volcanic bodies are comagmatic Riphean formations. Data on the age of the gabbro-norites are less definite. It is not impossible that these rocks are pre-Riphean, although their occurrence, petrographic composition, and mineralogical features suggest a genetic relationship with the western Bashkirian gabbro-diabases. The age of the Zubkovichi-Morino (Belorussia) gabbro rocks is not clear.

The early Riphean narrow troughlike downwarps appear to have originated in zones where the crystalline basement was disturbed. It is also clear that these very zones provided the most accessible vents for molten magmas on their way to the surface. Indeed, it has already been pointed out that volcanic activity was associated with the peripheral parts of grabenlike downwarps, with the zones of their junction with ancient shields. This is the situation along the east and north edges of the Voronezh massif, along the west, south, and east boundaries of the Moscow pre-Riphean shield. Rocks of southwestern Belorussia and western Bashkiriya are in a class by themselves. The structural position of these regions at the junction of deep downwarps predetermined a considerable shattering of the basement and a wide development of extrusives and vein rocks.

Volcanic activity was initiated on the east part of the Russian platform where the extrusives are observed among lower Bavly rocks (Lb). No volcanics have been found as yet

elsewhere on the platform in Serdobsk, Orsha, and Nenoksa rocks. The accumulation of Valday deposits (V) was accompanied by volcanic activity in the southwest part of the Russian platform and along edges of the Kresttsy, Kotlas, and Pachelma downwarps.

A Brief Petrographic Description of the Volcanics

A study of the Riphean volcanics on the Russian platform leads to their subdivision into three groups: intrusive, extrusive, and tuffaceous. These groups generally are so well differentiated that it is more convenient to consider them separately. In many instances, however, the division line between them is conditional, making it difficult to assign a rock to its proper group. This due, to a considerable extent, to the fact that the volcanics are covered with a sedimentary mantle, which precludes a detailed geologic study. Because of this, the proposed differentiation of (volcanic) rocks can be carried out only in a most general way and will have to be refined.

Petrographically, rocks of intrusive aspect belong to the gabbro series. They are the western Bashkirian gabbro-diabases, gabbro-norites of the eastern Tatariya and Volga region, and the biotite-pyroxene gabbro of Belorussia. Nearly all students of these formations note the uniformity of their mineralogic and chemical composition and structure. According to V.P. Florenskiy and T.A. Lapinskaya [23, 24], gabbro-diabases consist chiefly of labradoritic plagioclases -- andesine and monoclinic pyroxenes related to pigeonite and augite. They nearly always carry various amounts of quartz, commonly, in micropegmatitic growths with K-Na-feldspar, and with fairly common titanomagnetite, brown hornblende, and apatite. Olivine is present in some varieties. An instance of more basic rocks is the Or'yebashevo gabbro-diabases with a higher content of olivine pseudomorphs and with practically no micropegmatitic growths. Gabbro-diabases are generally coarse grained with a poorly developed plagioclase idiomorphism. A transitional doleritic texture is locally observed.

Gabbro-norites of the eastern Tatariya and Volga regions are close to gabbro-diabases in their composition, differing from the latter in their content of rhombic pyroxenes and hornblende. According to Yu. I. Polovinkina, the Zubkovichi-Morino magmatic body, in Belorussia, is made up of biotite-pyroxene gabbros and less commonly of gabbro-norites. These rocks carry various amounts of quartz, amphibolite, and limonite. Besides the gabbro-type formations, diorites occur at Zubkovichi, Morino, and Belitsa.

Extrusives are represented by assorted equivalents of the gabbro series. They are diabases, dolerites, and spilites along the slopes of the western Bashkirian grabenlike depressions; diabases along the east edge of the Volga-Ural massif; basalts, andesite-basalts, and diabases on the Voronezh massif's slopes; and finally, basalts, spilites, and trachydolerites along the edges of the Ratno-basement high (Beloressia). Under the microscope, spilites exhibit long and narrow bodies (leists) (0.8 to 1 mm) of albitized plagioclase with the space between them filled with chlorite which makes up as much as 40 percent of the rock volume. Subordinate quartz and K-feldspars form pegmatitic growths; also present are biotite, magnetite, and titanomagnetite. The rock structure is hyalo-ophitic.

Paleobasalts of the east edge of the Voronezh massif are marked by the very fresh aspect of their component minerals and by a high-silica content. A comparison of the western Bashkirian intrusive and extrusive rocks, as to their chemical composition, shows approximately the same oxide content in both. The only difference is the appreciably greater content of alkalies in paleodolerites and diabases.

Predominant among basic rocks of southwestern Belorussia are paleobasalts, with dolerites and noncrystallized lavas consisting of volcanic glass considerably less common. Three minerals are the main components of finely crystalline basic rocks: plagioclase of a labradorite or andesine type, augite, and magnetite, with the latter at times amounting to 10 percent. Rocks are commonly amygdaloid in structure, with the amygdules commonly filled with chlorite, chalcedony, zeolite, iron hydro-oxides, and other mineral aggregates. At times, they are left hollow. Basalts are, in many places, accompanied by spilites. They are usually greenish to red, with an abundance of amygdules. Spilites consist of irregularly oriented fibers (leists) of albitized plagioclases and fine crystals of ore minerals scattered in a strongly altered ground mass, which is not reactive to polarized light. Spilite amygdules are filled with the same substance as in basalts. Table 4 gives chemical analyses for extrusives.

Tuffaceous formations correspond to the basic series in their composition. They are represented by various clastic rocks: tuffs, tuffites, tuff breccias, and ash tuffs.

The southwestern Belorussian tuffs are chiefly vitroclastic, with vitrolithoclastics less common. Fragments are made of volcanic glass and a vesicular ferrous lava,

Table 4

Chemical Analysis of Extrusives on the Russian Platform
(After A.S. Makhnach [13], V.P. Florenskiy, and T.A. Lapinskaya [23, 24]

	Regions	Rocks	SiO_2	TiO_2	Al_2O_3	Fe_2O_3	FeO	MnO	CaO	MgO	P_2O_5	Na_2O	K_2O	SO_3	H_2O	Loss on ignition	Total
1	Dvino (B.S.S.R.)	Paleobasalt	47,95	1,49	15,35	10,09	6,28	—	8,90	7,20	0,18	0,93	0,19	Not determined	Not determined	1,31	99,87
2	Dvino (B.S.S.R.)	Paleobasalt	51,78	1,19	16,39	8,17	4,79	—	9,18	5,26	0,17	0,89	0,14	»	»	1,10	99,06
3	Malority (B.S.S.R.)	Lava	56,74	1,61	16,52	4,44	4,31	0,11	4,95	1,76	0,21	5,78	2,04	»	0,33	0,84	99,67
4	Dvino (B.S.S.R.)	Spilite	44,89	1,71	13,35	7,43	7,33	0,56	11,50	1,12	0,25	4,43	0,34	»	1,92	5,26	100,69
5	Malority (B.S.S.R.)	Spilite	51,79	1,83	10,23	7,84	9,20	0,43	4,09	6,32	0,23	3,99	0,53	»	0,98	2,97	100,13
6	Belorussia	Paleodolerite	48,75	2,12	14,53	7,50	4,40	0,02	8,93	8,37	0,37	1,98	0,36	None	5,00	2,45	99,79
7	Balashev	Paleobasalt	54,95	1,00	21,84	None	6,15	None	6,00	4,52	0,33	2,40	0,56	»	0,2	2,10	99,85
8	Staro-Petrovo (BashkASSR)	Paleodolerite	51,93	2,15	13,48	2,16	2,65	0,07	5,00	9,95	1,17	2,00	1,90	»	0,43	7,47	99,93
9	Staro-Petrovo (BashkASSR)	Paleodolerite	43,66	2,20	22,42	1,46	1,65	0,12	10,31	5,48	0,45	4,27	0,83	»	0,20	7,80	100,65
10	Suleyevo (Tatar A.S.S.R.)	Diabase	45,53	1,70	13,47	3,21	7,45	0,20	13,22	2,80	0,16	1,82	0,81	»	None	9,40	99,77
11	Suleyevo (Tatar A.S.S.R.)	Diabase	40,50	2,40	14,45	6,60	10,82	0,17	8,17	3,65	0,23	2,20	1,25	»	»	9,70	100,14

Note: comma represents decimal point

with basalt fragments less common. Clastic material is imbedded in a vitreous cement, strongly chloritized, ferrous, locally siliceous. In the Kaluga area, tuffaceous deposits are represented by lithoclastic and crystalloclastic formations, with the pyroclasts measuring from several millimeters to several centimeters. They consist of quartz grains and fragments of feldspar crystals, biotite, and of sedimentary and crystalline rocks in a cement of strongly chloritized volcanic glass. Some varieties contain as much as 75 percent by volume of volcanic glass fragments [5].

A place of their own is occupied by the tuff breccias of the Nenoksa area, west of Arkhangel'sk, into which one of the wells penetrated 110 m. In the neighboring boreholes, ancient sediments (N and V) were penetrated at the same stratigraphic level. It appears that the Nenoksa tuff breccias occur locally among sedimentary deposits. They are made up chiefly of nepheline basalt fragments [18].

In summing up the data on volcanic formations on various parts of the Russian platform, it may be stated that they all belong to a single petrographic type and that the variations within a rock do not take it out of its group. The affinity of the principal varieties of the gabbro-diabase series throughout the Russian platform, along with their areal extent, induced V.P. Florenskiy and T.A. Lapinskaya [23] to separate a diabase formation, correlative in their opinion with similar formations on other ancient platforms.

The definite tectonic position of Riphean volcanics makes it possible to predict their occurrence in places where they are either unknown or else suggested by indirect data, as it is along the southwest and southeast edges of the Russian platform and in its interior (Fig. 1). It should be noted, however, that the map does not show, by far, all the zones of the alleged manifestation of volcanic activity. It is very probable that the basement of the Moscow syneclise is not a single monolithic block but rather a series of differentially uplifted and depressed segments whose slopes were zones of magmatism. One of such zones may have passed through the Soligalich area where Devonian basalts and traces of sulfides in Carboniferous rocks have been found.

Volcanic activity on the Russian platform did not terminate with the Riphean. It revived in the Devonian — in Timan' and in the Bol'shoy Donbas downwarp (Chernigov, village of Mokraya Volnovakha). In addition, Devonian magmatic rocks have been found in the Radayevo depression, Volga-Ural massif (Kazaklar), in the central part of the Moscow

syneclise (Soligalich) and elsewhere. At the same time, monchiquite and camptonite-type rocks in the Carboniferous of the south edge of the Donbas show that Devonian volcanism, too, was not the last volcanic stage on the Russian platform. Aggregates of sulfide minerals which fill cavities and fissures in the ancient sedimentary complex, as well as in the younger deposits, suggest that volcanic phenomena on the Russian platform were followed by postmagmatic processes.

Further study will determine the tectonic position of Devonian and younger extrusive and vein formations, the structural types with which they occur, and the sequence in which various magmatic rocks appear at different stages in the history of the Russian platform. A solution of this problem will lead to a solution of similar problems in other ancient platforms and thus to general laws of volcanism on these most important structures of the earth's crust.

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Received May 5, 1958

STRATIGRAPHIC POSITION OF THE SINIAN COMPLEX[1]

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A lively controversy has been going on as to the correlation of Late Cambrian deposits which are identified by different names in different provinces, such as the Rocky Mountain Beltian, the Keweenawan of the North American platform, Norwegian sparagmite, Valday and Poles'y sequences of the Russian platform, the Uralian Riphean, the Vindhyan of India, the Australian Adelaide beds, Malmesbury beds of South Africa; post-Minas rocks of Brazil, etc. For this reason, the Interdepartmental Stratigraphic Committee was forced to adopt a compromise resolution by recognizing the Sinian as an intermediate between the Paleozoic and Proterozoic.

In the final reckoning, the problem is reduced to two arguments:

1. The Lower Cambrian is underlain by immediately preceding deposits. These deposits, be it only on some of the platforms, are in a conformable contact with the Cambrian, remain unmetamorphosed, and rest with a marked unconformity on older metamorphics of the platform base. Even in some folded provinces, the metamorphism of these deposits is not strong (as for instance the upper Precambrian of the Yenisey range, Baykal region, etc.). In such instances, as on platforms, they differ markedly from strongly altered ancient rocks and look more like Paleozoic rocks of the adjacent regions.

2. Paleontologically, upper Precambrian rocks differ greatly from the Cambrian. Although the highly organized and diversified Lower Cambrian organisms imply the existence of a fairly well-developed late Precambrian life, paleontologic traces of this life are virtually lacking in Sinian deposits, except for algae and spores. Thus, on paleontologic evidence, the Sinian complex differs most definitely from the overlying deposits.

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[1] K stratigraficheskomu polozheniyu siniyskogo kompleksa.

The first argument favors the association of the Sinian with the Paleozoic; the second, with the Proterozoic. There are, however, some additional considerations.

There is no doubt as to the Sinian participating, along with lower Paleozoic, in the development of many folded belts. It is almost certain that no geosynclines older than the Sinian are present in the Caledonian belts of Western Europe. This is suggested by the fact that deposits about a billion years old form the base of a platform fringing the folded belt. Some Lewisian gneisses of Scotland and Arendalion rocks in southern Scandinavia are of this nature. The Cordilleras belt is of the same age, with the earliest geosynclines of Beltian age.

The situation is different, however, in other places. There is every reason to believe that the so-called Riphean of the Urals includes rocks older than Sinian. This is specifically indicated by the suggested age for their granites, as much as a billion years. In the Baykal-Olekminsk region, L. I. Salop recognizes at least two pre-Paleozoic periods of folding. The upper limit for the first period is about 900 million years.

Thus, along with the Paleozoic, and for the same reasons, both the Sinian and older rocks may participate in a folding. Accordingly, the presence of Sinian rocks in a folded structure is not an argument for associating them with the Paleozoic. If this were not so, the older rocks, too, could be relegated to the Paleozoic. As witness the Baykal-Olekminsk region where the ancient complexes are as closely related to Sinian deposits as the Sinian are to the Caledonian of Western Europe.

Much more convincing is the argument that the Sinian on platforms is a part of the sedimentary mantle and is sharply divided from older metamorphics. This is the situation on the North American, Russian, Siberian, Chinese, and apparently the Tarim platforms. The Adelaide and Vindhyan complexes seem to corroborate that. These data are very

convincing and appear to favor a Paleozoic age for the Sinian.

However, these arguments are less persuasive if checked against all known platforms instead of a few selected ones. Indeed, the evidence is that the North American, Russian, Siberian, and Chinese platforms had been formed in a pre-Sinian time, and that Sinian deposits only formed a sedimentary mantle and did not participate in the basal structure. Furthermore, they are millions of years away from the platform base formation. Accordingly, Sinian deposits outside platforms participate in the formation of peripheral belts.

Some segments of the present platforms, however, may have been "consolidated" before the Sinian. Such segments may have persisted even into our time. Thus, deposits older than the Sinian may have participated in the building of their platform mantle. Such relationships are hardly probable among the small platforms of the northern hemisphere. They are more likely to occur among the large southern hemisphere platforms.

The recent argument on the age of the Jotnian and Ovruch sandstones of the Russian platform is well known. It was believed that their age as determined by the radioactive methods was too great. However, that very age was eventually accepted by the commission for absolute age determination, at the Academy of Sciences, U. S. S. R. The Jotnian and Ovruch have demonstrated that the Ukrainian and Scandinavian shields became platforms more than a billion years ago, and that their sedimentary mantle, nearly missing everywhere, is more than 1,100 to 1,200 million years old. The sediments, on both shields, are of course, closer in character to the Paleozoic than to the metamorphic complexes of the platform base.

Similar examples may be cited from Africa. The thick Transvaal sequence of South Africa, which was once tentatively assigned to the lower Paleozoic because of its slight metamorphism, is a typical platform mantle. Its accumulation is associated with the Katanga geosynclinal cycle. Thus, it was over 630 million years old at the close of the Katanga folding; in other words, the Transvaal sequence is much older than the Sinian.

The Witwatersrand sequence of the same region is also a part of platform mantle. It is much older than the Transvaal, probably assignable to the Kibarian folding (i.e. 1,200 to 1,400 million years), and its component uraninites date back 1,600 to 2,000 million years.

If the Sinian complex be assigned to the Paleozoic solely on its participation in the platform mantle, the same should be done for the Jotnian, Ovruch, Transvaal, and Witwatersrand. In that case, the lower boundary of the Paleozoic should be taken down to the oldest possible formations of the platform mantle. This is obviously absurd.

Let us turn now to the position of Sinian formations on other platforms. From what we know of the Gondwana, that immense, now disintegrated platform, included some very ancient segments. We mentioned some of them in connection with Transvaal and Witwatersrand rocks. However, the appearance of the platform as a whole is of a much later date. Whereas, the northern platforms already possessed their modern configurations, as early as the onset of the Sinian time or earlier, and were surrounded by still extant folded belts; (then young in their first or second folding period) ancient folded belts which were completing their development in the Late Cambrian of the Gondwana. Such are Late Cambrian structures of Minas Gerais and Rio de Janeiro, Brazil [6], the southwestern African folded structures [1, 3, 8], the Mozambique belt of A. Holmes [3, 8], and the Madagascar structures [10]. Of the same age are also the youngest movements and intrusions in Ceylon and Travancore, India [4]. Contemporaneous structures will probably be found in Australia and Antartica, where little has been done in the field of absolute-age determination. All these folded structures date back to the Sinian: about 500 million years for Brazil, 510 million years for southwestern Africa (Capetown granites), 480 to 490 million years for the Mozambique belt and Madagascar, and the same age for Ceylon.

These periods of folding followed more ancient ones, even as the Hercynian followed the Caledonian in Central Asia or Kazakhstan, or the Alpine the Hercynian in southern Europe. In Africa, terminal Precambrian was preceded by the Katangian, with the end of the latter's folding dating back 630 million years. In other words, late Precambrian folding belts had been initiated much earlier. By late Precambrian, they had a complex evolution behind them, which terminated at about the Cambrian boundary. Then the folded province was gradually transformed into a platform which, together with the adjacent older platform segments, formed a single great Gondwana platform. Thus, whereas the northern platforms date back at least 500 to 600 million years, the Gondwana is Paleozoic. It is at least 200 million years younger than the northern platforms.

This age difference between the Gondwana and northern platforms suggest a quite different position for the Sinian deposits on

them. In the north, Sinian and other Late Cambrian rocks either participate in the platform mantle or else form the early structures of still active "young" folded belts. In the Gondwana hemisphere, deposits of the same age either form a mantle on ancient platform segments, components of the Gondwana, or else participate in the structures of ancient, now inactive, folded belts, which are components of the platform bases of the southern hemisphere, along with ancient platform segments. Thus, the Sinian of the Gondwana platforms appears to be older than the platform as a whole. It participates in its metamorphic base and is more closely related to Precambrian than to Paleozoic deposits.

We now see that the structural argument for a Paleozoic age of the Sinian is not as simple as it seemed. Its apparent logic is based chiefly on the fact that we know something, at least, of the northern hemisphere geology and practically nothing of the "Gondwana hemisphere." If the reasoning of those who defend a Paleozoic age for the Sinian complex be applied to the Gondwana data, a structural affinity of this complex with the Proterozoic is undeniable and so is its divorce from the Paleozoic.

Changing now to the paleontologic aspect of the age of the Sinian complex, the break between the Cambrian and even the uppermost Precambrian is well known. Now, as against the diversified and fairly common Lower Cambrian fauna, the Sinian complex contains but very rare and always somewhat doubtful fossil remains. This means that the paleontologic method of age determination - effective from Lower Cambrian on - is difficult and unreliable for Sinian deposits. In any event, it does not permit any detailed differentiation of the Sinian. In this case, it is based only on algae and spores. It has been established that spores are but crude age criteria; as to algae, they cannot determine the age even down to a system. As a result, the paleontologic method is not as applicable to Sinian rocks as it is for younger formations.

Is it possible that, despite this great difference, Sinian deposits are paleontologically closer to Paleozoic than to older sediments? Should the Sinian be regarded as an epoch of spores and lower algae, characterized by such forms as stromatolites? Is it possible that lower beds are generally devoid of organic remains? If so, this would have been the strongest argument for assigning the Sinian complex to the Paleozoic.

As a matter of fact, however, there is no sharp break between the Sinian and Proterozoic. Algal formations from lower Paleozoic and Sinian deposits are also known from more ancient rocks. The Katanga stromatolites are virtually the same as the Sinian, although Katanga deposits and the contemporaneous sequences were folded and intruded prior to the development of Sinian geosynclines. If the close of the Sinian is dated about 500 million years ago, with the beginning of it perhaps 600 million years, terminal intrusions of the Katanga cycle must be 615 to 650 million years old. Then, the accumulation of sediments in geosynclines should be assigned to a period which started no less than 700 million years ago.

According to B. V. Timofayev, spores are known to occur in Precambrian time, i. e. at at epoch corresponding to at least the Katangian. Thus, they too cannot be accepted as Sinian index fossils.

If the finding of primitive algae and spores were a basis for assigning their carrier rocks to the Paleozoic, the base of the latter would have to be moved far down, to include the Katangian. Furthermore, there are data, albeit isolated, pointing to a much older age of similar algae. Such are for instance the findings of structures thought to be algae in limestones of the oldest known South African sedimentary sequence which comprises the Sebakwian, Bulawayan, and Shamvaian groups [1]. The extremely old age of this sequence is determined not only stratigraphically but by the absolute-age method as well: granites which cut it are dated as 2,650 million years old. Of course, organic remains are preserved only in exceptional instances in such old rocks. On the other hand, such findings, if confirmed, show that there is no reason to assign to the upper Precambrian (Sinian) all algae-carrying Precambrian rocks. It appears that the development of these organisms, or rather the alteration of their end structures, proceeded very slowly. This does not make them a good dating means.

Thus, we again see that the paleontologic method, which works well from the Cambrian on, gets us into serious difficulties if applied to the Sinian. It is hardly helpful in any differentiation of these deposits, let alone the older ones. Another method of dating ancient formations is needed. This is found in radioactive decay. This method as a rule, seems to be inferior to the paleontologic, for Paleozoic and younger rocks. But the reverse is true for Precambrian where the paleontologic method, even if applicable, is so inexact as to be ineffective, thus making the absolute-age method the only constant guide to the history of the earth.

It is to be noted that the absolute-age determination method does not have the drawback of other dating methods. Like the paleontologic method, it is not tied up with processes and phenomena to be dated. This

is the reason that absolute-age determination, and it alone, is a substitute for the paleontologic method which is ineffective for ancient rocks.

What is the main point of this brief discussion? First of all, it is the conditional character of "natural boundaries" between the Paleozoic and the pre-Paleozoic, which of course is true for all stratigraphic boundaries. This boundary should not be drawn on structural data even if only because the boundaries do not exactly coincide chronologically. Thus, the upper boundary of the Baykalian, Beltian, and later African folded bodies is chronologically appreciably different. This is the reason that the age should be determined irrespective of structural relationships. We have seen, in addition, that both the upper and lower Sinian boundaries have different meanings in different parts of the world. As a result, the main argument for a Paleozoic age of the Sinian, which is based on the association of this sequence with the mantle of northern platforms, is invalid. On southern platforms, the Sinian sequence is involved in the structure of their bases. Consequently, the position of this sequence as a whole does not favor its reference to either the Paleozoic or Proterozoic.

Paleontologically, the Sinian sequence differs most radically from the Cambrian and not too much, generally, from underlying formations, provided of course that the latter are not metamorphosed. On this basis, it hardly can be assigned to the Paleozoic.

It must be admitted, on the whole, that the arguments of those who attempt to draw stratigraphic boundaries chiefly on the basis of structural relationship are unconvincing, even disregarding the inadmissible practice of begging the question by measuring with what is to be measured. Such arguments should be disregarded. From a practical point of view, it is more expedient to separate the Sinian from the Paleozoic. This would be in line with their distinct paleontologic boundary and with the fact that, from the Sinian down, the absolute-age method must be substituted for the paleontologic method. It might be advisable to separate the Sinian and underlying deposits, up to about 700 to 800 million years of age, as a major subdivision, perhaps reserving for it the familiar name of upper Proterozoic. The precarious position, assigned to the Sinian by the Stratigraphic Committee, can hardly be called a happy choice. According to it, the underlying sequence - closely related to it, especially by the character of its organic remains - also should be assigned to the Proterozoic. As any compromise solution, the resolution on the position of the position of the Sinian deposits is hardly realistic.

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Received September 30, 1957

GLACIAL-MARINE DEPOSITS IN THE YENISEY REGION, WEST SIBERIAN PLAIN[1]

by

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Glacial-marine deposits were first identified in the U. S. S. R. by N. A. Kulik [6], in 1915, for the middle and lower Pechora basin. It appears, however, that he greatly exaggerated the extent of these deposits. Commenting on his data, I. I. Krasnov [5] states, "Marine moraines possibly do exist, but their distribution in maritime regions is small." Never-the less, S. A. Yakovlev [12], too, notes a moraine on the Pechora plain, containing "whole shells of Astarte borealis Chemn, Saxicava arctica L., Mya truncata L., Tellina calcarea Chemn." It appears, then, that glacial-marine sediments are not as restricted in that area as is stated by I. I. Krasnov.

Similar deposits of various ages have been noted in many places throughout the north of the West Siberian plain.

In 1941, V. N. Saks [9] described moraine-like loams with boulders and a deep water fauna of Portlandia lenticula Möll., P. frigida Torel., Arca glacialis Gray, Neaera arctica M.S., etc, from the lower Yenisey area. He tentatively correlated them with the Central Siberian plateau and the Taymyr glaciation epoch. However, he retracted his correlation as early as 1945 [10]. Nowhere on the Central Siberian plateau, the Byrranga Mountains, and in their foothills, did he find any trace of the Sanchugov glaciation which should have been reflected in a regular increase of the boulder and pebble content in the Sanchugov layer, with the approach to those highlands, and in the transition of marine sediments to the moraine. The enrichment of marine deposits in clastic material, to the extent that they "assume the aspect of marine moraines," he ascribes to marine erosion of the terminal moraine and to river load.

However, V. N. Saks' assumptions are in direct contradiction with field data of the Noril'sk geologists. In 1953, G. D. Maslov stated unequivocally that "the amount of clastic material above the Messovsk sands (i. e. in the Sanchugov layer - S. A. A.) increases toward the Central Siberian plateau." The well-sorted, fauna-carrying sediments disappear in the same direction and "the deposits assume a morainal aspect." Maslov rejects V. N. Saks' ideas on the manner of entry of clastic material into the Sanchugov basin; he notes the haphazard distribution of boulders and pebbles in marine sediments as compared to their concentration in a basal layer and their separation from finer argillaceous material, as should be the case in a transgression. In a river deposit, too, pebble beds would have been laid in bands and isolated patches, which is here not the case.

G. D. Maslov emphasizes further that the terminal moraine in the Dudinka area near Yenisey, along the Fokina, Kosaya, Dudinka and other rivers, is covered with 30 to 80 m of Messovsk sands and could not be eroded by the Sanchugov transgression.

Furthermore, despite V. N. Saks' [11] assertions of an alluvial origin of the Messovsk sediments, the Noril'sk geologists established a marine origin. This was done in 1953, first by A. V. Kulikov, then by S. L. Troitskiy for the lower Yenisey course; and by S. B. Shatskiy, in 1956, for the Pura basin (Samburgsk borehole). This means that a Sanchugov sea could not have eroded the terminal moraine, and that the sea itself was a further development of the Messovsk-Samburgsk transgression. Deposits of these basins are connected by gradual transitions, with basal pebble beds, conglomerates, and boulder sands (local remains of the Samarov terminal moraine) lying at the base of the Messovsk and Samburgsk sequences.

The Noril'sk geologists also doubt the basic assertion of V. N. Saks on the absence of traces of a glaciation in mountains of the West Siberian plateau, contemporaneous with the Sanchugov transgression. In 1953, G. D. Maslov stated that "Field data show that a

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[1] K voprosu o sushchestvovanii glyatsial'nomorskikh Otlozheniy v priyeniseyskom rayone zapadno-sibirskoy nizmennosti.

Sanchugov glaciation indeed took place on the Siberian platform. It was a valley type and apparently occurred sporadically." In this connection, A. V. Kulikov noted in 1953 a wide development of glacial-marine deposits at the edge of the Noril'sk Mountains. These deposits antedate the second (i.e. Zyryansk) glaciation.

These data induced N. N. Urvantsev to state, in 1954, that "glaciation in the Noril'sk region did not cease with the boreal transgression, although it was substantially restricted. Both in the beginning and at the end of the transgression, glaciers were active enough to unload their ice and clastic load directly into the sea."

Independently of the Noril'sk geologists, glacial-marine deposits throughout the eastern half of the West Siberian plain were identified by A. I. Popov [15], in 1948; and by others [1, 2, 3, 4, 10], in 1953-1957.

In 1953-1956, the Tazov glacial bed was separated as a stage of a maximum [7] or else an independent glaciation [4]. It was subsequently demonstrated that this bed from the middle Yenisey basin and adjacent areas is stratigraphically correlative with the Sanchugov beds identified by V. N. Saks along the lower Yenisey and at Ust'-Port [1, 2, 4, 7]. Direct lithofacies observations incontrovertibly point to a lateral transition of the Tazov deposits to the Sanchugov, from south to north; with a continuous facies series, from the subaerial Tazov moraine at the edge of the Central Siberian plateau to glacial-moraine deposits of the Yenisey depression and to the predominately marine deposits, along the lower Yenisey.

At the plateau edge, in the headwaters of the Fokina River and Didinka River (as observed by Maslov, in 1953) and south from there, along the Devyatikha, Tatarka, Fat'yanikha and other rivers, the subaerial Tazov moraine is represented by gray and ash-gray sandy loams and loams containing a detrital structure and a haphazard arrangement of boulders, pebbles, and gravel. Screen analysis (Table 1) shows 20 to 23 percent clastic material consisting of comparatively numerous boulders (to 9.1 percent) and coarse pebbles (to 6.3 percent). Clay minerals, as in the Samarov moraine, are represented by hydromicas.

The Tazov moraine rests with an erosional unconformity upon alluvial-lacustrine deposits. It is in no way different externally from the Samarov moraine with which it merges into a single glacial sequence along the mountain slopes. However, the Tazov moraine assumes a progressive glacial-marine aspect, going north along the Yenisey in its right-bank cliffs. Both gradual and abrupt changes from coarse sand and sandy loam deposits with boulders, pebbles, and gravel, to clay oozes and typically lenticular clays are observed laterally and in sections. There are local intercalations of dustlike thin-bedded sands (Bakhtinskiy Yar). Clastic material is either stratified or scattered haphazardly throughout the section. This is best seen at the Alinskiy Yar where the upper part of the outcropping section exhibits ash-gray loams, with a greenish, locally bluish cast, somewhat lumpy and almost devoid of coarse clastic material. They gradually change downward to typical boulder loams with lentils of clays and, in places, of sands.

Farther down the Yenisey, in the Markovskiy, Chernoostrovskiy, and Pupkovskiy Yars, also in the lower course of Turukha River and in the Tolstyy Nos area on the Nizhnyaya Baikha River the glacial-marine sequence becomes lenticular and stratified, with a sharp differentiation into boulder and boulder-free deposits. As an example, the Markovskiy Yar section is as follows, reading downward:

1. Soil, 0.20 m.

2. Sandy loam, straw-colored, dustlike, cryptostratified, gradually changing to the underlying loams; 1.2 to 1.5 m.

3. Loam, brown-gray, sandy, stratified, the stratification brought about by thin (3 to 5 mm) partings of sandy loam and dustlike sand; lumpy, with a few fine pebbles and abundant gravel; 1.5 to 1.8 m.

4. Sand, quartz, rusty-yellow, stratified, with lenses of gravel and few pebbles; 0.4 to 0.5 m.

5. Sand, quartz, gray with brown tint, fine grained, thin to very thin bedded; 0.3 to 0.4 m.

6. Loam, gray-brown, clayey, splintery-lumpy, with angular pebbles, small boulders and gravel; 1.2 to 1.3 m.

7. Loam, similar to 6 but with alternation of thin drab-brown clay and dustlike sandy loam; 0.2 to 0.3 m.

8. Sandy loam, yellow, dustlike, changing to clayey sand with wavy very thin stratification; 0.2 to 0.3 m.

9. Loam similar to 6, cryptostratified, with angular boulders, pebbles, and gravel; 0.3 to 0.35 m.

10. Loam, dark gray with brownish tint, sandy, massive, full of splintery rubble, gravel, pebbles, and boulders; 0.5 m.

Table 1

Granulometric Composition of the Tazov-Sanchugov Deposits
(Grain size in mm)

Location	Boulders	Pebbles			Gravel		Sand			Dust and clay	Porosity
	100	—100+50	—50+25	—25+10	−10+5	—5+2	—2+1	—1+0,5	—0,5+0,25	0,25	
Along the west edge of the Central Siberian Plateau											
Right bank of Kosma River, 1-2 km above Cheremshanka brook	9,1	5,8	1,3	1,78	0,26	0,95	0,40	0,10	—	71,0	8,32
Right bank of Fat'yanikha River, 1-2 km above the Bezymyannyy brook	7.1	6,3	2,8	0,78	1,25	1,25	0,2	—	—	73,02	6,3
Right bank of Tatarka River, 0.3 km above the Polunochnyy brook	5,5	3,3	4,4	3,9	2,2	0,55	0,83	0,05	—	69,87	9,4
Along the right bank of the Yenisey											
Bakhtinskiy Yar	1,2	5,7	0,9	1,0	0,90	1,13	0,9	1,13	1,36	80,55	6,13
Bakhtinskiy Yar	—	2,9	1,10	1,80	0,55	1,21	Traces	Traces	Traces	84,44	8,0
Alinskiy Yar	—	—	0,9	2,1	3,65	2,25	0,80	1,25	1,65	79,0	8,4
Kangotovskiy Cape	5,5	2,4	1,8	1,1	0,81	1,30	0,55	0,8	2,70	74,64	8,4
Markovskiy Yar	1,5	3,45	1,15	2,85	5,60	2,35	5,13	1,67	4,50	66,0	5,8
Chernoostrovskiy Yar	—	1,25	1,25	3,12	2,85	3,25	1,0	Traces	Traces	81,08	6,2
Chernoostrovskiy Yar	—	—	2,50	6,90	6,87	13,65	10,0	»	»	52,88	7,8
Pupkovskiy Yar	—	—	2,25	3,6	0,6	1,0	1,25	»	»	81,6	9,7
In the Yenisey depression											
Tolstyy Nos, 1.5 km above mouth of Merkheka River	—	—	—	2,3	1,65	2,15	0,6	0,25	Traces	84,35	8,7
Along the lower Yenisey											
Right bank of Fokina River, 7 km above the Tokureyev brook	—	2,0	0,6	1,4	1,4	2,0	1,4	2,0	0,6	88,9	9,7
Right bank of Yenisey River, northern edge of Dudinka	—	—	—	4,75	2,60	5,35	1,30	Traces	Traces	78,50	7,5

Note: comma represents decimal point

11. Loam, boulderfree, thin bedded, clayey with 3-5 mm partings of dark gray clay and fine-grained dustlike sand; 0.3 to 0.5 m.

12. Loam similar to 10, with coarse angular boulders; 1.5 to 1.6 m.

13. Gravelly pebble bed, 0.2 to 0.25 m.

14. Quartz sand, yellow-gray, fine grained, cross bedded; 0.2 to 0.3 m.

15. Loam, similar to 10; 0.3 to 0.4 m.

16. Sand, brown-yellow, mixed grained, stratified; 0.3 to 0.4 m.

17. Loam, similar to 10; 0.2 m.

18. Sand, similar to 16; 0.5 to 0.6 m.

19. Loam, similar to 10; 9.0 to 9.5 m.

20. Loam, ash-gray, sandy, splintery-lumpy, with boulders, pebbles, and abundant gravel; 1.0 m.

Below that, there are the Messovsk-Samburgsk sands.

In the summer of 1956, V.A. Zubakov and D.V. Semevskiy [3] found fragments and shells of marine molluscs in marine-glacial deposits, both in situ and on the beach along the above-mentioned cliffs, as distant as the latitude of the mouth of the Yeloguy River. V.A. Zubakov lists [3, 4] Arctic species of Portlandia arctica Gray, Astarte crenata (Gray), etc, and also abundant Saxicava arctica L., Macoma baltica L., suggesting the freshening of the basin. The best undamaged specimens in situ were found in the Pupkovo area. In the summer of 1956, the author collected whole Portlandia arctica and P. lenticula Möll. shells (identified by S.L. Troitskiy) from the Tazov-Sanchugov bed, in the Tolstyy Nos area (Nizhnyaya Baikha River) at about the same latitude. A marine fauna in the Tazov moraine was discovered recently by Yu. A. Lavrushin [1, 2] from the Turukhan Basin, and identified by M.A. Lavrova as Macoma sp. (calcarea?) L. and Astarte crenata Gray. Thus, in situ marine fauna has been positively identified in the Tazov glacial layer 200 to 250 km south of the Sanchugov-Basin boundary, according to V.N. Saks [17].

It is not the fauna alone which points to a glacial-marine origin of the Tazov layer in this area. There are parallel and very significant lithofacies changes. Instead of a typical moraine, unsorted and rich in clastic material (Table 1) along the plateau, outcrops along the Yenisey and even boreholes along the Turukhan disclose glacial-marine sediments, lenticularly stratified and obviously sorted and differentiated. They consist of lentils of boulder loams and sandy loams, fine grained to dustlike sands, and gravel; with stratification ranging from coarse cross bedding to extremely thin horizontal; fine oozelike clays, loams, etc. An appreciably lower overall clastic content is observed, along with a decrease in its grain size, until boulders and coarse pebbles completely disappear (Table 1). Traces of montmorillonite and beidellite appear alongside the hydromicas (Chernoostrovskiy and Pupkovskiy Yars). Of interest are glauconite grains found by N.G. Zaikina (oral communication) in the Pupkovskiy Yar sediments.

Farther down the Yenisey, stratified marine clays gain ascendency over lenticular coarse loams with pebbles and gravel, as exposed in glacial-marine outcrops in the Plakhino area, along the Fokina River; between the Uboynaya and Bol'shaya Avamskaya Rivers. Here, boulders and coarse pebbles are lacking, as a rule; instead, there are small pebbles, 25 ± 10 mm, and gravel. The content of clay particles <0.001 m rises considerably, with clay minerals represented by a mixture of hydromicas and beidellite, the latter predominating in some specimens. The clays are also more abundant in fauna.

The following were identified by S.L. Troitskiy in the Plakhino area, in the right-bank cliffs of the Yenisey, as distant as the Fokina River: Portlandia arctica (Gray), Astarte cf. montagni (Dill.), A. compressa (Linne), Arca sp., Saxicava arctica (Linne), etc. The "classical" Sanchugov fauna with Portlandia lenticula (Möll.), Arca glacialis Gray, also Portlandia arctica (Gray), Nucula tenuis (Mont.), Astarte compressa (Linne), Macoma calcarea (Chemn.) first appears along the Fokina.

The Tazov-Sanchugov layer facies terminates in the Dudinka area and farther down the Yenisey, where a very uniform sequence of dark gray to bluish loams and clays is exposed. They are massive to well stratified, detrital, usually nearly free of clastic material. The latter is accumulated only locally, disturbing somewhat the uniformity and the high degree of sorting of the sediments. According to V.N. Saks, these deposits carry a deep and cold water fauna of Portlandia lenticula Moll., Arca glacialis Gray, and other arctic forms, with boreal and subarctic species lacking.

Thus, the fauna assemblage south of the Fokina River, and especially south of Igarka and Turukhansk, is considerably poorer than that of Ust'-Port. At the same time, there is evidence of the freshening of the basin (presence of Saxicava arctica L., Macoma baltica

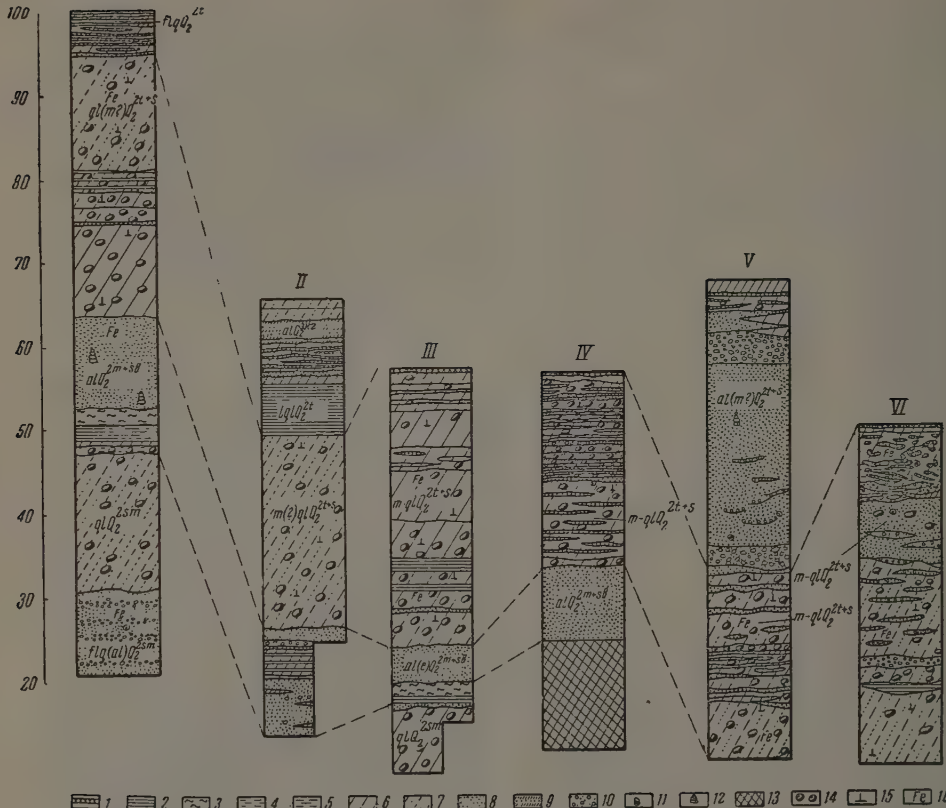

FIGURE 1. Typical sections of middle Pleistocene glacial, glacial-marine, and alluvial deposits along the right bank of the Yenisey, between Bakhta and Turukhansk.

I -- Bakhtinskiy Yar; II -- Kangotovskiy Yar; III -- southern edge of Alinskoye; IV -- Markovskiy Yar; V -- Chernoostrovskiy Yar; VI -- Pupkovskiy Yar.

Lithology: 1 -- soil; 2 -- lenticular clay; 3 -- oozelike clay; 4 -- sandy clay; 5 -- sticky clay; 6 -- loam; 7 -- sandy loam; 8 -- sand; 9 -- alternation of sand and sandy loams; 10 -- pebble bed; 11 -- marine mollusc fauna; 12 -- fresh water fauna; 13 -- sodded slopes; 14 -- boulder-pebble-gravel inclusions; 15 -- carbonate traces; 16 -- ferrous traces.

Stratigraphy: alQ_2^{kz} -- alluvial deposits of above-flood-plain terrace III of the Yenisey; $fglQ_2^{2t+s}$ -- fluvioglacial deposits of the Tasov glaciation stage; $fglQ_2^{2t+s}$ -- lacustrine-glacial banded clays of the Tazov maximum glaciation stage; $al(m?)Q_2^{2t+s}$ -- alluvial (littoral-marine deltaic?) deposits of above-flood-plain terrace IV; gl(m) and $m\text{-}glQ_2^{2t+s}$ -- glacial-marine (water) deposits of the Tazov maximum glaciation stage (Tazov-Sanchugov layer); $al(l)Q_2^{2m+s^b}$ -- alluvial (lacustrine) Messovsk-Samburgsk sands and clays; glQ_2^{2sm} -- moraine of the Samarovsk maximum glaciation stage; $fgl(al)Q_2^{2sm}$ -- fluvio-glacial alluvial pebble beds and sand of the Samarovsk maximum glaciation stage.

L., Portlandia arctica (Gray, etc).

All this fairly substantially confirms the presence of glacial-marine deposits in the north of Western Siberia and the Russian platform. This is of great scientific and practical significance in relation to one of the fundamental problems of Quaternary geology: the correlation of marine and continental deposits.

Data on glacial-marine deposits of the Yenisey area, show that the relationship between transgressions and glaciations is considerably more complicated than it appears from the glacial-eustatic hypothesis [13]. According to the latter, Quaternary transgressions developed as a result of ice melting, with regressions caused by its accumulation on continents.

Without denying the existence of eustatic changes in the level of a world ocean, a certain abstract quality of these speculations should be noted. First, it is difficult to imagine a mechanism of a fully equivalent transition from sea water to continental ice, and vice versa. With warming, evaporation will increase along with the degradation of glaciers; whereas, in cooling, the glaciers will grow (partly) at the expense of atmospheric moisture. This very complicated water cycle should, in our opinion, considerably weaken the eustatic effect.

Secondly, one cannot ignore the effect of tectonic movements which may radically alter the results of eustatic shore-line oscillation. Indeed, there are convincing data gathered from tectonic activity, during historical times, in various parts of Western Siberia, the Soviet Arctica, etc., [11].

Commenting on the data on hand, V.N. Saks concludes that "tectonic movements played the principal part in Quaternary shoreline oscillations. Eustatic and isostatic factors were only superimposed on these movements, bringing about their increase or decrease, amounting at times to their change of sign."

Finally, it would be quite erroneous to ignore the reverse effect of transgressions on glaciations, even granting the latter a leading part in the course of events during recent times. The increased humidity, unavoidably resulting from a transgression, might give an impetus to the development of glacial phenomena in markedly continental Siberia.

Transgressions, raising the erosion base, penetrated deep into the continent and contributed to the onset of lacustrine conditions in south-central parts of Western Siberia. It appears that the greatest penetration occurred in pre-Samarovsk time, when isolated large lakes may have been connected with the sea [11]. This also is the most humid period of the Siberian Quaternary. It must be emphasized that the lower Pleistocene transgression in the Soviet Arctic unquestionably antedated the earliest glacial phenomena in the mountains [11]. As such, it could not have been a result of the melting of continental ice.

Hardly fortuitous is the coincidence of the numbers of transgressions and glaciations in Western Siberia, and the definite relationship of their magnitude. Finally, it is very significant that in Eastern Siberia, where the transgressions were small, the magnitude of glaciations, as well as their number (?), were considerably lower. However, the positive effect of transgressions on the development of glaciers was manifest only under proper climatic conditions - on the background of cooling periods. A transgression, contemporaneous with the apparently general warming of the northern hemisphere, played a negative part, by contributing to the obliteration of glaciers in the mountains.

In the light of the above, the presence of glacial-marine deposits appears to be a direct and very graphic instance of the reverse effect of transgressions on the development of glaciations. Seen in this light, the formation of the above-described deposits is fully rational. Glacial-marine deposits originate from sizable deviations from eustatic oscillations, as a result of tectonic, and especially differential tectonic, movements. The latter undoubtedly took place in Soviet Arctica [11] as mentioned above.

It may be conjectured, in this connection, that glacial-marine conditions of sedimentation prevailed in different regions in different times. As a proof, there is the maximum Samarovsk glaciation synchronous with the Salemal' transgression; it is established for the lower Ob', by G.I. Lazukov from voluminous and convincing material.

Thus, we believe that glacial-marine deposits result from very complex interaction of transgressions and glaciations, on a favorable background of differential tectonic movements and paleoclimate.

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Received October 25, 1957

MICACEOUS PEGMATITES AND THE ABSOLUTE AGE OF POST-JURASSIC INTRUSIONS IN THE ALDAN [1]

by

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Pegmatites, hitherto unknown among younger intrusives of the Aldan, are described, with data on the absolute-age determination by the K-Ar-method. Geologic observations suggest an Upper Jurassic age for most of this magmatic complex.

* * * * *

A complex body of alkaline-earth syenites and alkaline rocks, concentrated in horizontal Cambrian dolomites and Jurassic fresh-water sandstones of the Aldan shield, upper structural stage has the earmarks of a hypabyssal origin. More than fifty petrographic varieties, described earlier from this complex [3, 4, 5], have a porphyritic, porphyrylike, or uniformly medium-grained texture. Typical vein formations, associated with alkaline syenites, are represented by fairly common veins of fine-grained syenite-aplites that are more like granites.

Pegmatitelike coarse-grained rocks were observed only within the Yakutsk complex intrusion, 17 to 20 km south of Aldan (Fig. 1). This massif, 6.5 x 3.5 km in area, is represented at the surface by bald summits: the Lebedinyy, Yakut, Pereval'nyy, and others. Its southern and southeastern parts are made up of variegated quasicontemporaneous alkaline rocks: eruptive breccias of alkaline orthophyres, pseudoleucitic porphyries, nepheline and muscovite syenites. Alkaline rocks in the north and northwest parts of the massif are intruded by augite syenites varying in composition from monzonites to syenites.

Pegmatites occur in both alkaline earth and alkaline rocks. A belt of quartz-orthoclase fragments, evidently a trace of a small vein, was observed in eluvium of augite syenites, on top of the Lebedinyy summit. Individual quartz-orthoclase growths attain eight centimeters. The microscope reveals typical pegmatite growths of quartz and of strongly pelitized orthoclase ($-2V=73^{o}$). A more interesting variety is observed in muscovite syenites in the southeast part of the massif, on the right-hand slope of the Pionerskiy spring. Here, a prospector's quarry uncovers a biotite-anorthoclase pegmatite, no less than 15 meters wide, whose position, details, and relationship with the enclosing muscovite-syenites, are not clear.

The internal structure of this rock is characterized by irregular growths of biotite in a macroscopically monomineral matrix of gray anorthoclase with tabular crystals, from 1 to 10 cm in diameter. A peculiar feature of the biotite distribution is the predominance of large crystals -- usually more than 5 to 10 cm and as much as 50 cm -- in some growths (Fig. 2), whereas other growths are made up of small scales (about 1 mm). Under the microscope, the fine-scaled biotite exhibits pleochroism, from greenish-brown. The refraction indexes are β = 1.625; γ = 1.620; α = 1.571. Biotite swells slightly, on heating, although its overall water content is normal for unaltered mica (Penfield pipe determination, 1.43%; Berg pipe, 1.92%).

Under the microscope, anorthosite ($-2V = 65^{o}$) exhibits a very small amount of decay perthitization, and of pelitization with minute solid particles, at times showing red color. Apparently, this is hematite, formed in decomposition of the solid solution. Very fine sericite scales occur in some coarse anorthoclase crystals and sericite plates (0.1-0.2 mm) in fine-grained aggregates form thin, broken microveinlets with negligible amounts of green biotite, magnetite, and chlorite.

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[1] Slyudonosnyye Pegmatity I Absolyutnyy Vozrast Posleyurskikh Intruziy Aldana.

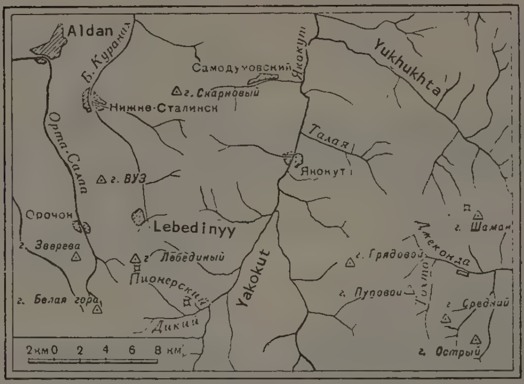

FIGURE 1. Index map of Central Aldan. Triangles mark localities of specimens for the absolute-age determination.

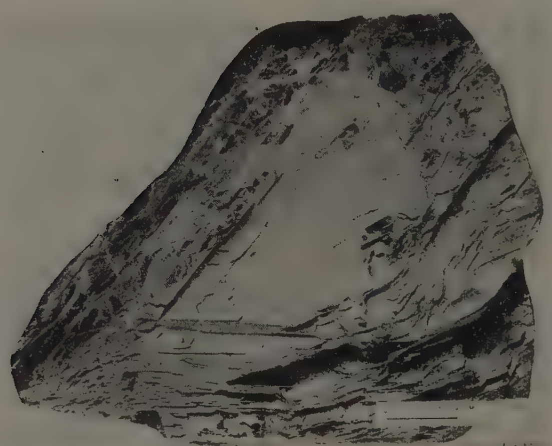

FIGURE 2. A coarse biotite crystal in biotite-anorthosite pegmatite. Dikiy Spring basin.

The presence of such coarse-grained rocks in a definitely hypabyssal extrusive complex is rather surprising, especially considering that they belong to pegmatites of Type One and early geophases of Fersman, which are supposed to be the deepest seated. However, the classical hypabyssal complex at Crestmore (California), assigned on the basis of its mineral and structural features to the shallowest subvolcanic formations, also contains coarse pegmatites with crystals as much as 30 cm, cutting intrusive porphyritic rocks and skarns [9]. Further study is necessary to explain the origin of such hypabyssal pegmatites.

Inasmuch as there are no previous determinations of the absolute age of post-Jurassic rocks of the Aldan, it was of interest to use the coarse monocrystals of feldspar and biotite from the pegmatite for this purpose.

The only other materials, suitable for this purpose, were K-feldspar incrustations in aegirine-augite alkaline syenite (laurvikite) from the Shaman summit (upper course of the Dzhekonda). This is a large laccolith (3 x 1.2 km) associated with the Archaean-

Cambrian contact. The laurvikite consists chiefly of K-Na-feldspar, more or less perthitized; a variable amount of aegirine-augite, very little albite and some hornblende or brown biotite; for accessory minerals, there are apatite, sphene, and magnetite. Its texture is medium to fine grained, locally with trachytoid tabular feldspar. Some enlarged feldspar crystals (as long as 5 cm) are observed. Such porphyritic bodies locally literally fill the standard fine grained matrix.

Under the microscope, the feldspar crystal used for the absolute age determination, exhibited several small (hundredths and tenths of mm) idiomorphic inclusions of aegirine-augite, apatite, sphene, and magnetite. The feldspar was unevenly perthitized. Its isolated segments, virtually devoid of perthite growth and pelite, give an impression of relicts in a predominately perthite groundmass. The amount of growth in perthite is also variable. Tiny fibers of albite, in places, account for not more than 10 percent of the volume; but it also forms coarser (as much as 0.1 mm) irregularly tabular bodies with definite polysynthetic twins, with the amount of albite reaching 30 to 40 percent.

Judging by the amount of alkalis (K_2O - 7.42%; Na_2O - 6.58%; M.G. Zamuruyeva, analyst) this crystal is 43.90 percent orthoclase and 55.78 percent albite. Since the amount of albite growth does not exceed 15 percent, K-Na-feldspar should contain no less than 30 percent albite. Inasmuch as the optical axes' angle of this feldspar, as determined on Fedorov's table, is 84°, it may be called sodium-rich orthoclase.

Specimen 900-53 was studied in the Laboratory of Precambrian Geology at the Academy of Sciences, U.S.S.R. (E.K. Gerling). The remaining specimens were studied by S.S. Sardarov at the Laboratory of the Dagestan Affiliate of the Academy.

Discrepancies in the absolute age as determined in the Precambrian Laboratory and in the Dagestan Affiliate, from contemporaneous pegmatite biotites, exceed only slightly (5 to 6%) the average divergence limit for these methods, experimentally determined [1]. Inasmuch as the cause of the lesser age for biotite is not clear, and since the absolute-age values for contemporaneous mica and feldspar usually are in inverse ratio, the absolute age obtained for feldspars from various intrusions should be accepted as the true one. The figures thus obtained correspond to the Upper Jurassic of our geochronological column.

These figures, together with some geologic observations, make it possible to date most of the rocks of this intrusive body.

Yu. A. Bilibin believed that the majority of younger intrusives in the Aldan belong to a single post-Jurassic magmatic cycle. Later, I.V. Belov [2] voiced a belief of their pre-Jurassic age, because of the lack of earlier sills in the Jurassic. Findings of trachytoid-rock pebbles similar to eruptive breccia from the Ostryy summit alkaline orthophyre (upper course of the Dzhekonda) also suggested to him their pre-Jurassic age. However, this author traced two sills of hornblende porphyries, altering Jurassic sandstones, on the Pupov summit. The close similarity of these sills to two other sills in the nearby Cambrian dolomites suggests that they are of the same age. On the other hand, Jurassic sandstones are cut by eruptive breccia of alkaline orthophyre on the south slope of the Ostryy summit. These observations favor the earlier ideas of Yu. A. Bilibin who also found porphyritic fragments in the basal Jurassic but thought them alien to the area [3]. Thus, the lower age limit of this extrusive body is determined by the age of the Jurassic deposits carrying an abundant Bajocian-Bathonian flora and fresh-water fauna [7].

Our study of the interrelationship of these intrusions suggests that the age sequence of the four major groups of Yu. A. Bilibin is still valid, save for a few details. The laurvikites belong to the third age group, with the pegmatites cutting its syenites. The figures for the absolute age of these rocks define, then, the upper-age limit for most

Data of the absolute-age determination

Mineral	Specimen No.	Content			$\frac{mAr}{mK^{40}}$	Age in million years
		K (% weight)	K^{40} g/g	Ar cm^3/g		
Biotite	900--53	7.76	$9.31 \cdot 10^{-6}$	$3.49 \cdot 10^{-5}$	0.0067	107
Biotite	911--53	7.55	$9.06 \cdot 10^{-6}$	$3.70 \cdot 10^{-5}$	0.0072	120
Orthoclase	907--53	10.50	$12.60 \cdot 10^{-6}$	$5.70 \cdot 10^{-5}$	0.0081	135
Anorthoclase	197-b--53	7.42	$8.90 \cdot 10^{-6}$	$4.00 \cdot 10^{-5}$	0.0081	135

intrusions, with the exception of the very restricted aegirine granites of the fourth age group.

This concept of the age limit for post-Jurassic intrusions refines the former ideas on the subject. In determining the age of the intrusions, Yu. A. Bilibin [3] used indirect evidence. Because of the drainage-divide position of the laccoliths, he considered their intrusion to be not later than the Tertiary. Yu. K. Dzevanovskiy [6] postulated a post-Cretaceous age for the younger Aldan intrusions by extrapolating their position from the east edge of the Aldan shield. Such a far-reaching extrapolation in a tectonically different area can hardly be convincing. The drainage-divide position of the laccoliths of Bilibin's argument may be explained, not by their recent intrusion, but by a recent general uplift and by their resistance to erosion. The latter hypothesis finds some justification in the Ye. B. Pavlovskiy [8] concept of recent uplifts in the Aldan.

Thus, all these data lead to a conclusion that the entire young Aldan intrusive body came into being in a comparatively short geologic period, namely in the Upper-Jurassic epoch. Despite this pinpointing of their age, we believe that these intrusions may retain their conventional name of "post-Jurassic," since they are younger than all Jurassic deposits of the region.

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Received October 25, 1957

EVIDENCE OF MESOZOIC-CENOZOIC VOLCANISM ON THE NORTHERN EDGE OF THE SIBERIAN PLATFORM[1]

by

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In the course of the 1954 geologic and geomorphologic prospecting along the middle course of the Popigay River, a right tributary of the Khatanga, the author studied the widely developed volcanic rocks represented by agglomeratic to ash tuffs, tuff breccias, tuff lavas, and in places, lavas.

A Permian-Triassic age for these formations has been agreed upon, since 1948 (D.V. Kozhevin). L.P. Smirnov [6], who mapped the Popigay depression on a scale of 1:1,000,000, in 1951, assigned them conditionally to the Triassic (T?). There was no doubt as to their belonging to a trap formation, despite D.V. Kozhevin's finding of diabase dikes cutting the Permian-Triassic tuffs, and of Jurassic wood in some tuffs. This gave him a chance to postulate a different age for the tuffs; nevertheless, on his map, all of the Popigay volcanics were lumped as Permian-Triassic.

In 1954, this author observed a fauna-carrying calcareous sand concretion, among other inclusions in tuff breccia along the Daldyn' River. Numerous fragments of carbonized and silicified wood and lignite were also collected. The lignite was in no way different from Cretaceous brown coals outcropping in the area. These findings led the author to surmise a younger than Cretaceous age for some of the pyroclastics developed in the Popigay depression (Fig. 1).

In the laboratory processing of the 1954 collection, N.I. Shul'gina (Scientific Research Institute for Arctic Geology) identified the fauna as Pseudomonotis (Eumorphotis) (?) cf. lenaensis Lah, which is an index fossil for the Aalenian (Middle Jurassic) in the Popigay basin.

The specimen was poorly preserved and its identification was tentative. However, the probability of Jurassic fragments in tuff breccias was corroborated indirectly by the results of study of fossil wood, abundant in tuffs and in a number of outcrops. According to I.A. Shilkina (Botanical Institute, Academy of Sciences, U.S.S.R.), the wood remains belong to the genus Xenoxylon, namely to Xenoxylon meysterii Palib. et Jarm.

According to A.V. Yarmolenko, this genus is indicative of the Jurassic, whereas Xenoxylon meysterii is an index fossil for the Jurassic of Central Asia, the Trans-Baykal, Far East, and Siberia. Thus, the presence of Jurassic rocks in the tuff breccias is well substantiated, and the study of spore-pollen assemblages from brown Popigay coals and their gangue rocks has established an Albian-Cenomanian age for the coals (as determined by N.M. Bondarenko, Scientific Research Institute of Arctic Geology). Considering the time necessary for burial and carbonization of wood, it may be assumed that volcanic activity in the Popigay depression was initiated considerably later than the Albian-Cenomanian; perhaps as late as the Tertiary.

In their chemical composition, these allegedly Tertiary volcanics turned out to be more acid than similar rocks of the Permian-Triassic trap formation, whose fragments were found among other inclusions in the material of younger tuffs and tuff breccias. Given below are the results of a chemical analysis of the vitreous cement fraction of Tertiary (?) tuff breccia, by the chemical laboratory of the Lensoviet Technological Institute. The specimens (Fig. 2 and 3) were taken from the main outcrop on the Chordu-Daldyn River (exposure 206). Exposed there is a massive volcanic sequence, full of fragments and larger pieces of diverse rocks whose age ranges from the Archaean to and including the Upper Cretaceous.

The largest of these pieces are more than a meter in diameter and are, as a rule, of granite-carrying Archaean crystalline schists

[1] O Proyavlenii Mezokaynozoyskogo Vulkanizma Na Severnoy Okraine Sibirskoy Platformy.

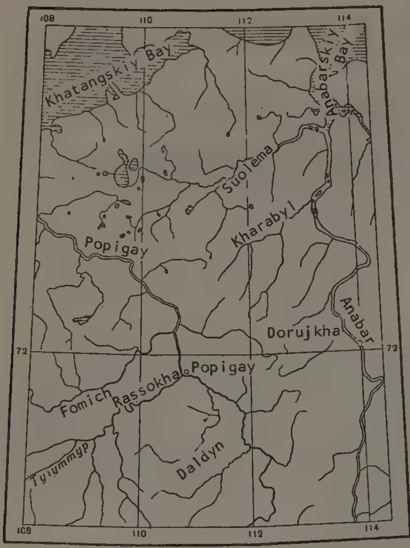

FIGURE 1. Index map of the Popigay depression.

with less common Valanginian sandstones. The Archaean fragments are always rounded, with a burnt, light-colored surface. The sandstone fragments are angular and exhibit no obvious traces of baking or melting. The total amount of coarse clastic (more than 50 cm) material reaches 20 to 25 percent. The bulk of the tuff breccia is a friable rock full of fragments, from fractions of a centimeter to a decimeter. Locally it is saturated with glass and is, therefore, more dense. The finest grained variety presents a dense mass with earthy fracture. The rock abounds in amygdules filled with secondary minerals, and with conspicuous curved inclusions of dark lustrous glass.

Under the microscope, clastic material in the specimen under study is seen to amount to 65 to 72 percent, with predominating fragments of glass (48 to 50 percent) and quartz (10 to 12 percent). Feldspars (microcline and plagioclase) account for 2 to 3 percent. Also present are fragments of biotite, granite, rhombic and monoclinic pyroxene (less than 1 percent), and assorted rocks (5 to 6 percent). Because of the uneven distribution of fragments in tuff breccia, their true relationship with the cementing mass is impossible to determine from thin sections.

The groundmass of cement consists of glass, chlorite, and fine-grained brown mica. Segments of a vitreous mass of average basicity (n for glass = 1, 552) alternate with those of ash.

The inclusion glass is fresh, unaltered, evenly colored pinkish-lavender, locally nearly colorless. It is even more acid than the groundmass glass (n = 1, 537 $\pm$ 0, 004, which corresponds to a SiO_2 content as much as 58 percent). A characteristic feature of the inclusion glass is its perlitic and flow structures.

For a more precise analysis of the chemical composition of Tertiary (?) lava, its results were corrected to the amount of foreign inclusions represented, as mentioned above, chiefly by quartz with an admixture of K-feldspar and acid plagioclase. To be sure, even with this sum subtracted, the results obtained should be regarded as highly

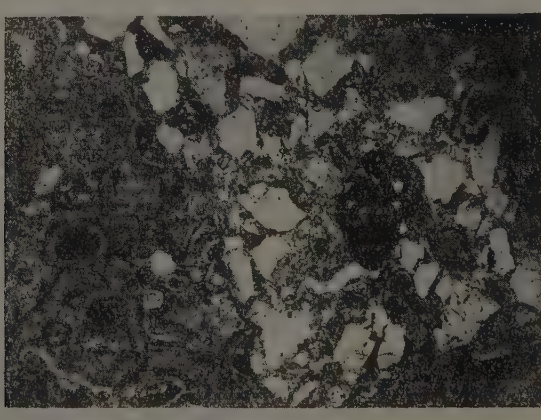

FIGURE 2. Andesite-basalt tuff lava from cement of the Mesozoic-Cenozoic tuff breccia. Thin section 206-a; without analyzer; 44 X.

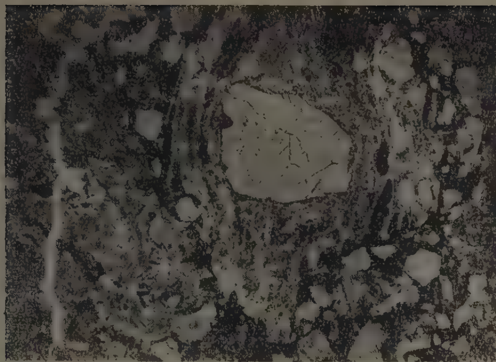

FIGURE 3. Flow structure about a quartz inclusion, in a median-composition glass, within a litho-vitro-crystalloclastic tuff from cement of a Mesozoic-Cenozoic tuff breccia. Thin section 206; without analyzer; 44 X.

approximate, inasmuch as the amount of inclusions of such rocks is variable and the method of determination of modifying factors, not quite perfect.

As shown by the analysis, Tertiary (?) lavas differ from normal Permian-Triassic traps of Siberia by their greater silica content. In their composition, they are close to augite andesite (R. Daly) but are less alkaline.

All of this points to a Mesozoic-Cenozoic volcanic cycle at the north edge of the Siberian platform, apparently not related to trap volcanism. Indeed, the lower boundary of the latter does not usually rise above the Lower Triassic, with a Middle Triassic-Lower Jurassic interval possible only for its post-lava intrusive phase (5).

The importance of explosive rocks was established in 1954. They are a result of a predominately explosive volcanic activity localized in a comparatively small area along the Popigay middle course. Considering the development features of this platform segment, very peculiar tectonically, the existing Popigay hollow, known in the geological literature as a quasi-graben [6], is a volcano-tectonic depression.[1] It originated in a pre-Cretaceous time, at the site of a collapsed domelike uplift, with outlines different from the present. Ever since, this area has been very mobile, keeping up its rising tendency. Most evident are the manifestations of Upper Cretaceous volcanism, apparently a result of the opening of fissures and of a fast influx of lava, also of post-Cretaceous -- most probably Tertiary and perhaps Quaternary -- dislocations. The latter resulted in a series of radial and concentric normal faults. It is possible that they were accompanied by explosions. It is significant that these Mesozoic-Cenozoic volcanic explosions consisted chiefly in eruptions of explosive material with small amounts of intermediate lava. Conditions for a quiet and protracted flow of lavas apparently did not prevail.

This concept of the origin of this hollow, as well as of the Mesozoic-Cenozoic age of pyroclastics developed in its area, first had but a little support among the students of the north part of the Siberian platform.

In 1956, areal geologic surveying on a scale of 1:200,000 and prospecting for useful minerals were initiated in the northwest sector of the Popigay hollow. New data were obtained, confirming the Mesozoic-Cenozoic age of local volcanics and their wider than anticipated distribution. These data (by L.P. Smirnov, Ye. I. Podkopayev, and others), together with the results of the study of fauna, spore-pollen assemblages and fossil wood, established the occurrence of major post-Aptian and pre-Neogene eruptions chiefly of an explosive type, in the west part of the hollow. L.P. Smirnov, demonstrated that the explosions alternated with periods of relative quiescence with normal alluvium, at which time lacustrine sediments and tuffaceous material were laid down in valleys and lakes.

Yet, geologists who have worked in the Popigay hollow note that indigenous tuffs and lava tuffs possess certain features which

[1]P.S. Voronov (1954) regards the hollow as a "sinking crater."

complicate their stratification. It was noted for instance, that beds cannot be traced over any extended distance (Ye. I. Podkopayev, 1957). Even with a single exposure it is often difficult to fix the boundaries of different rocks. The transition from one variety to another is accomplished through a change in the amount of different inclusions, the intensity of baking of ash, and the degree of its saturation with glass. The glass is so unevenly distributed in clastic mass that a thin section from the same field specimen may be identified either as a vitro-litho-crystalloclastic, or ash tuff, or a tuff lava, or even a lava.

L. P. Smirnov, I. A. Sidorchuk, and Ye. I. Podkopayev are inclined to see in the Popigay tuff lavas and lavas those peculiar rocks, ignimbrites, which are formed in the Katmai-type explosions, through baking and partial recrystallization of incandescent volcanics consisting of minute particles of glass, sand, rock fragments, and a semiliquid lava. Without getting into a discussion of such an origin of ignimbrites [2], their similarity to the Popigay tuff lavas and "lavas" cannot be denied. This similarity is as follows:

1. Rocks separated as "lavas" and tuff lavas by the students of the Popigay hollow do

Chemical composition of Tertiary (?) tuff breccia cement, from 1954 collection.

Component	% weight			
	specimen 206-a		augite-andesite (after R. Daly)	average composition of the Tunguska basin trap (after A. P. Lebedev)
	tuff breccia cement: glass and small inclusions	cement glass, without small inclusions		
SiO_2	65,22	58,5	57,50	48,50
TiO_2	0,69	0,1	0,79	1,42
Al_2O_3	12,59	15,4	17,33	15,75
Fe_2O_3	3,20	4,1	3,78	3,43
FeO	3,34	4,2	3,62	8,88
MnO	0,08	0,1	0,22	0,19
MgO	2,62	3,3	2,86	5,62
CaO	4,58	5,6	5,83	10,69
Na_2O	1,11	1,2	3,53	2,18
K_2O	2,16	2,2	2,36	0,69
P_2O_5	—	—	0,30	—
Cr_2O_3	0,21	0,2	—	—
H_2O^+	—	—	–	1,27
H_2O^-	—	—	—	1,38
Loss on ignition	4,00	5,1	1,88	—
Total	99,80	100,00	100,00	100,00
Of that, moisture content	2,59			

Numerical characteristics, after A. N. Zavaritskiy

	Specimen 206-a (groundmass without inclusions)	Augite andesite (after R. Daly)	Average composition of traps (after A. P. Lebedev)
a	6,0	11,6	5,9
c	7,3	6,3	8,0
b	15,4	13,4	27,5
s	71,3	68,7	58,6
Q	+23,3	+ 7,9	— 2,6

not form flows in the proper meaning of this term. They rather are transitionally related to agglomerates and ash tuffs. The "lavas" always carry certain amounts of inclusions of foreign minerals and rocks. The dimensions of these inclusions from fractions of a millimeter to 2 cm, are such as to preclude their sinking into an acid, and consequently a fairly viscous lava, under the action of gravity alone. It is more correct to regard these rocks only as tuff lavas.

2. Despite the sizable thickness (as much as 100 m) of tuff lavas and tuffs, the glass in their central parts is very poorly crystallized, with the crystals formed chiefly about foreign inclusions, which results in a comb structure. There are segments of baked ash and of incipient spherulite crystallization.

3. Like ignimbrites and true lavas, the Popigay tuff lavas are marked by a well-developed columnar parting. Like ignimbrites, they are characterized by an uneven composition, with all the gradations from tuffs to tuff lavas and even lavas, which form small but distinct segments in the tuffs.

A complete similarity between the Popigay rocks and ignimbrites is hardly to be expected, however, first of all because of the difference in their chemical composition. The Katmai ignimbrites and tuff for lavas of Armenia are related to acid magmas. On the other hand, the presence of a labradorite type plagioclase (Nos. 54-56) and augite, in microinclusions, along with acid glass (n from 1.537 ± 0.004 to 1.501 ± 0.003) in the Popigay lavas (data by M.T. Kiryushina, 1955; and L.P. Smirnov, 1957) renders their acid composition very doubtful. Most likely, they are close to augite andesites, i.e., to derivatives of a basalt magma.

Furthermore, the Armenian ignimbrites, studied by A.N. Zavaritskiy [3], consist chiefly of finely pulverized ash and lava, and of laminated lava inclusions ejected in small amounts along with incandescent clouds of sand and ash. Coarser material is of a subordinate significance in the Valley of a Thousand Smokes [3]. This relationship is reversed in the Popigay hollow. Here, agglomeratic tuffs and tuff breccias commonly predominate, and the entire body of deposits is more like a Pelée type of eruption with its explosive paroxysm collapsing the vault of the magmatic hearth, ejecting immense amounts of explosive material, and resulting in a well-defined relief of volcano-tectonic hollows. As a rule, there are no lava flows associated with such explosive eruptions [1, 2, 5].

With regard to the saturation of tuffs with glass and to the presence of flow structures, fairly common in the Popigay tuff lavas, these phenomena may be explained by a secondary penetration of lava into the overlying pyroclastics shattered by the explosions and perhaps somewhat cooled, by that time. These rocks require detailed study and perhaps rate a special name which would reflect their origin and would prevent their confusion with ignimbrites.

All this emphasizes the tectonic factor in the origin of the present Popigay hollow which is regarded as a purely erosional feature by some of its students.

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Received October 21, 1957

HYDROCARBON GASES AND BITUMENS IN INTRUSIVE MASSIFS OF THE CENTRAL KOLA PENINSULA[1]

by

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Prior to 1951, no one suspected that rocks of the Khibinskiy nepheline syenite massif had a high combustible gas content. According to B.M. Melent'yev, who studied these gases in 1951 and 1954, the Yukspor tunnel gas had the following composition: methane and its homologs, 81.4 percent; hydrogen, 13.9 percent; oxygen, 0.4 percent; nitrogen and other inert gases, 5.2 percent.

In 1956-1957, the author conducted a study of gas content in the Khibinskiy and Lovozero alkaline massifs and in the Monchegory massif of basic and ultrabasic rocks. The Khibinskiy massif was studied in great detail as to the gas in microfractures and permeable pores, and as to bitumens in extrusive rocks. Areas of gas distribution throughout the Khibinskiy tundra were outlined by geochemical methods, with drilling of holes, 1 to 1.5 m deep. Only the pore-space gas and bitumens were studied in the Lovozero and Monchegory massifs.

The Khibinskiy nepheline syenite massif presents a complex intrusive body, stratified, located in the central part of the Kola peninsula.[2] To the north and southeast, it contacts Archean gneisses; in the south and west, it is surrounded by the Imandra-Varzuga formation of greenstones and tuffaceous-sedimentary rocks, which is assigned to the Proterozoic.

The peripheral and central parts of the massif are made up of coarse-grained nepheline syenite: khibinite and foyaite. Arranged in circular arcs between them, there is a series of mixed-grained nepheline syenites and iolite-urtite rocks with apatite-nepheline intrusions.

Apatite-nepheline rocks are associated with the hanging wall of the iolite-urtite exposures and are conformable with them. The apatite content in the apatite-nepheline body increases upward from its base.

The Lovozero massif, like the Khibinskiy, is a complex intrusive body of several intrusive phases. In contrast to the Khibinskiy massif, which is of a central intrusion type, it is an inclined tabular plutonic body underlain by Archean gneisses.

The Monchegory basic to ultrabasic massif, like the other two, is associated with a tectonic zone along the Archean-Proterozoic contact. It has two structural stages: the lower stage is represented by Archean gneisses; the upper, by sedimentary, tuffaceous, and extrusive Proterozoic rocks. The intrusive body is located between these stages and is made up of ultrabasic to basic rocks: pyroxenites, peridotites, gabbro-norites, and norites.

Geochemical study has revealed that higher methane concentrations in subsoil deposits are associated with the Khibinskiy massif and are lacking outside of it, in the area of development of the Imandra-Varzuga sedimentary-extrusive formation, and of Archean gneisses.

The study of gases in microfractures and interconnected pores (free phase) of the Khibinskiy massif was made in apatite-nepheline mines. It, in all instances, demonstrated the presence of hydrocarbons in these rocks.

Side-wall hole tests in mines have a low hydrocarbon content for gases, not more than 0.01 to 0.001 percent, as a rule. On this background, there are segments of higher hydrocarbon gas content, as much as 18 percent. Hydrogen is less common and occurs in lower concentrations, not more than 0.7 percent.

The hydrocarbon fraction of gases is represented usually by methane with considerably

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[1] Uglevodorodnyye Gazy i bitumy intruzivnykh massivov tsentral'noy chasti kol'skogo poluostrova.

[2] See geologic map in A. A. Polkanov's paper [5].

smaller amounts of ethane, propane, and butane.

Segments of stronger gas shows are associated with rocks of the iolite-urtite sequence shich underlies the ore body, and with comparatively apatite-poor lenticular apatite-nepheline rocks.

The maximum concentration of hydrocarbon gases is associated with urtites.

As determined by side-wall tests, the compositions of the free-gas phase is as follows (percent of volume):

O_2	15—20.9	H_2	0—0.7
CO_2	0—1.97	N_2	65.31—81
CO	0	He	0.0011—0.0013
CH_4	0.00028—16.5	Ar	0.96—0.98
C_2H_6	0.00022—1.40	$\frac{Ar \cdot 100}{N_2 \cdot 1.18}$	=1.03—1.08
C_3H_8	0.00011—0.23		
C_4H_{10}	0—0.156		

It appears that the fracture and pore gas in the Khibinskiy rocks is a mixture of a deep-seated gas and air, which invades them through numerous fractures. The extraction of gas from pores was done by pulverizing the specimens in an airtight ball mill, down to 0.1 to 0.01 mm.

Gases in closed-rock pores were studied throughout the Khibinskiy massif section, and to some extent in the Imandra-Varzuga sedimentary-extrusive formation. In 1957, reconnaissance work was carried out in the Lovozero and Monchegory intrusive massifs. Specimens were collected from mines, core holes, and outcrops. Most attention was given to the study of the Khibinskiy rocks.

It was found that combustible gases from closed pores of the Khibinskiy tundra rocks embrace the entire range of saturated hydrocarbons: methane, ethane, propane, butane and, in smaller amounts, hydrogen. Rocks differ sharply in hydrocarbon gas content (Fig. 1). For the average gas content, see Table 1.

For the nitrogen-content determination, the desorption was carried out for three specimens after the air in the ball mill had been replaced by CO_2. It was found that the pore gas contained a considerable amount of nitrogen, as follows (gas content in c.c. for 1 kg of khibinite):

Specimen number	Hydrocarbon gases	H_2	CO	N_2
223	22.87	1.08	0	156.0
225	14.28	4.08	0	177.90
227	27.42	2.44	0	56.44

These data show that sedimentary extrusive rocks of the Imandra-Varzuga formation are marked by a low methane content and by the lack of heavy gaseous hydrocarbons C_2—C_4

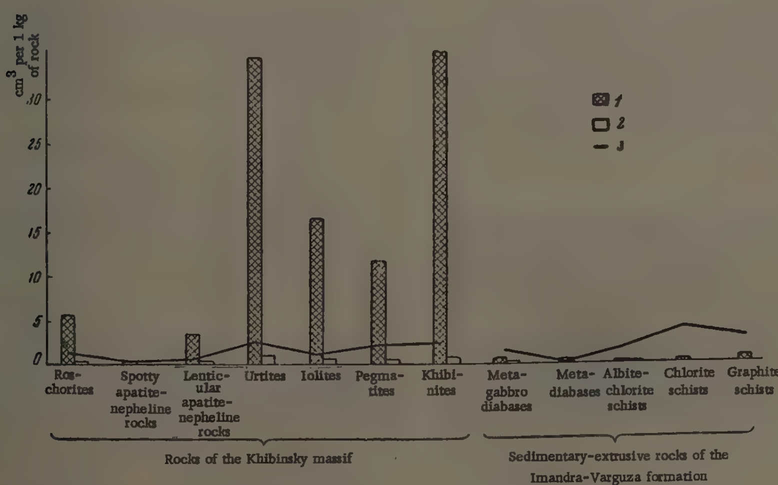

FIGURE 1. Content of combustible gases in rocks of the Khibinskiy tundra. 1 -- methane; 2 -- heavy hydrocarbons; 3 -- hydrogen

Table 1

Average Gas Content in Closed Pores of Khibinskiy Tundra Rocks and of the Imandra-Varuga Formation

		CH_4	C_2H_6	C_3H_8	C_4H_{10}	H_2	CO	CO_2	
I. Khibinskiy massif rocks									
Foyaites	8	17,95	0,05	0,03	0,005	0,58	2,17	—	20,5
Roscherite	8	5,42	0,04	0,06	0,03	1,14	0,47	14,03	15,4 —17
Spotty apatite-nephelines	3	0,06	0,05	0	0	0,09	0,93	12,2	6,7 —8
Lenticular apatite-nephelines	15	3,29	0,10	0,10	0,02	0,43	0,63	6,75	14,4 —15,6
Urtites	8	34,77	0,45	0,44	0	2,46	0,48	16,06	23,57—25,37
Iolites	10	16,65	0,20	0,27	0	1,12	0,95	25,10	18,5 —20
Pegmatites	9	11,78	0,18	0,21	0,07	2,38	0,77	15,02	—
Khibinites	5	46,51	0,26	0,52	0	2,57	0,60	17,0	21,5 —22
II. Sedimentary-extrusive rocks of the Imandra-Varuga formation									
Metagabbro-diabases	2	0,53	0,08	0	0	1,49	4,28	52,00	—
Metadiabases	1	0,45	0	0	0	0	—	—	—
Albite-chlorite schists	2	0,06	0,01	0	0	1,59	0,07	3,45	—
Chlorite schists	3	0,14	0	0	0	4,11	0,04	0	—
Graphite schists	2	0,71	0	0	0	3,62	0,12	5,10	—

Note: comma represents decimal point.

Intrusive rocks of the Khibinskiy massif carry a sizable amount of hydrocarbon gases in their closed pores (as much as 50 c.c. to 1 kg of rock). Heavy hydrocarbons were noted in all specimens, along with a fairly high concentration of CO_2. The hydrogen content in different rocks varies from 0.09 to 9.13 c.c. to one kg, with its higher content usually corresponding to a higher hydrocarbon gas content.

A comparison of data obtained by different methods shows that the results of gas analysis from surface outcrops, from side-wall tests, and from closed pores are in good agreement with each other.

The relationship between the amount of hydrocarbon gases in closed pores of an intrusive rock and the chemical composition of the latter is significant.

In pores of the Khibinskiy plutonic rocks, the hydrocarbon content is proportional to their aluminum content (Fig. 2).

Eighteen specimens from the Lovozero nepheline-syenite massif were studied. They were taken from mines and core holes throughout practically the entire plutonic interval, from the underlying gneisses to Devonian tuffaceous schists. The gas content in closed pores is given in Table 2. Specimens were not analyzed for nitrogen.

Gneisses are shown to contain small amounts of hydrocarbon gases (0.58 c.c.) and hydrogen (0.46 c.c.) and a comparatively high concentration of carbon dioxide (12.2 c.c.). Hydrogen is present in combustible gases to an amount not exceeding hundredths of one cubic centimeter.

Nepheline syenites of the massif are characterized by a comparatively high hydrocarbon-gas content, although not as high as in Khibinskiy. Their highest concentrations are associated with foyalites, 4.47 c.c., and eudyalite-carrying urtite of layer one, 5.11 percent. Besides the predominating methane, hydrocarbon gases contain C_2–C_4.

As against the Khibinskiy massif, the Lovozero nepheline syenites carry practically no carbon dioxide; traces were found only in two specimens. All specimens, except

Table 2

Analysis of Gases in Closed Pores of the Lovozero Massif Rocks

	CH_4	C_2H_6	C_3H_8	C_4H_{10}	H_2	CO	CO_2
Gneisses	0,477	0,061	0	0,075	0,46	0,23	12,2
Foyaites	4,28	0,065	0,131	0	0,77	0,11	2,74
Loparite louyavarites underlying the layer-5 utrites	1,845	0,053	0,043	0,03	0,13	0	—
Eudyalitic urtites of later 1	3,487	0,691	0,226	0,71	0,9	0	3,66
Foyalites with filliaumite	0,578	0,035	0,014	0,021	0,28	0	0
Ore-bearing foyalites with villiaumite inclusions	0,543	0,068	0,017	0,034	0,19	0	0
Malignites	1,12	0,085	0	0	0,62	0,24	—
Ore-bearing urtites with villiaumite inclusions	0,94	0,14	0,047	0	0,17	0	0
Urtites with some villiaumite inclusions	0,161	0	0	0	0,20	0	0
Louyavrites, porphyritic	0,033	0,028	0	0	0	—	—
Louyavrites, eudyalitic	2,72	0,34	0,084	0,10	0,75	0	0
Tuffaceous schists	0,637	0,053	0,053	0	28,35	0,52	10,50

Table 3

Average Gas Content in Closed Pores of the Monchegory Massif Rocks

Rocks	Gas content, in c.c. to 1 kg of rock						
	CH_4	C_2H_6	C_3H_8	C_4H_{10}	H_2	CO	CO_2
Gneisses	0,12	0,007	0,007	0,010	1,83	0,33	18,25
Gabbro-norites	0,05	0	0	0	1,79	0,7	15,16
Norites	0,03	0	0	0	1,55	0,34	32,40
Peridotites	0,13	0	0	0	1,45	1,79	25,32
Pyrrhotite ore	0,19	0	0	0	0,30	0,60	24,22
Pyroxenites	0,03	0	0	0	0	0	13,5
Olivine pyroxenites	0,09	0	0	0	0,22	4,34	18,97

Note: Comma represents decimal point.

porphyritic louyavrites, show a low hydrogen content (0.17 to 0.9 c.c.). Devonian tuffaceous schists are marked by a high hydrogen content, 28.35 percent, with a considerable concentration of carbon dioxide.

Seventeen specimens from the Monchegory plutonic section were studied, ranging from the underlying gneisses to pyroxenites. In the amount and composition of their pore gas, Monchegory gneisses are similar to the Lovozero.

In their hydrocarbon-gas content, basic and ultrabasic intrusive rocks differ sharply from alkaline rocks of the Lovozero and Khibinskiy massifs. All of the samples show a lack of of heavy hydrocarbon gases, whereas the methane content in the pores is very low, mostly in hundredths of a c.c. and not exceeding 0.19 c.c. to 1 kg of rock. The hydrogen-content decreases from 1.79 to 0.22 c.c., going up the section; hydrogen was missing in one specimen (Table 3).

The finding of hydrocarbon gases in extrusive rocks stimulated interest in their bitumen content. The results of a chemical-bituminous analysis of the Khibinskiy tundra rocks is given in Table 4.

The content of the Khibinskiy bitumen chloroform extract differs sharply from that of crudes, in its low content of oils: 14.98

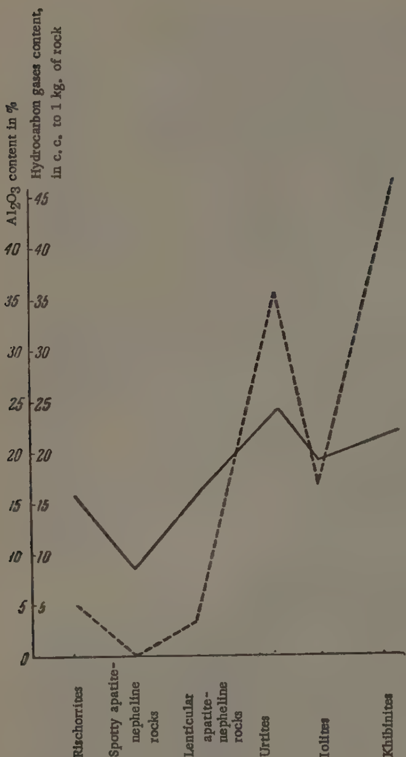

FIGURE 2. Comparison of hydrocarbon-gas content (broken line) in closed pores and aluminum content (solid line) in the Khibinskiy massif rocks

Table 4

Chemical-bitumen Analysis of Khibinskiy Rocks

Rocks	Chloroform extract content, in % rock volume	Component content: Oils	Benzol tars	Alcohol-benzol tars	Asphalts	Elementary composition of chloroform extract: C	H	N+O+S	Elementary composition of fractions: C	H	N+O+S	Fractions	Amount of analyzed rock
Lenticular apatite-nepheline	0.00053	—	—	—	—	77.71	12.35	9.94	—	—	—		1.8
Same	0.0004	14.98	43.24	34.81	6.00	79.80	11.63	N 0.13 — S 0.47	82.96	10.26	6.78	Oils	20.0
									77.45	9.31	13.24	Benzol tars	
									69.05	8.93	22.02	Alcohol-benzol tars	
Iolite-urtite	0.0006	—	—	—	—	81.02	13.55	S 0.19	—	—	—		5.18
Khibinite	0.0013	—	—	—	—	83.37	13.28	S 0.86	—	—	—	—	3.92

percent as compared to the usual 50 to 80 percent oil content in crudes. There is a marked decrease in benzol tars (43.24 percent) as compared to the alcohol-benzol (34.81 percent) in the Khibinskiy bitumen, which suggests a reduced character. In its elementary composition, this bitumen is close to that of crudes and to the scattered bitumens in sedimentary rocks of oil areas. The carbon content in crudes usually varies from 80 to 88 percent; of hydrogen, from 10 to 14 percent; and the C:H ratio is 5.6 to 9. In our case, C:H = 6 to 6.8; the carbon content is 77.7 to 83.37 percent; hydrogen 11.63 to 13.55 percent.

An interesting relationship in the Khibinskiy rocks is observed between the amounts of

their bitumens and their hydrocarbon gases in closed pores. The former increase with the latter, and the elementary composition of such more abundant bitumens approaches that of bitumens in crudes (Khibinites: C = 88.37 percent and H = 13.28 percent; lenticular apatite-nepheline rocks: C from 77.7 to 77.9 percent; H from 11.63 to 12.35 percent).

A few of the Lovozero specimens were analyzed for bitumens by the luminescent method. Small amounts of oily and tar bitumens, from 0.00015 to 0.0012, were detected.

Only traces of bitumens have been found, on the whole, in the Monchegory-massif rocks, with only 4 out of 16 specimens showing a bitumen content of the order of 0.00015 to 0.003.

SUMMARY

1. Alkaline extrusives of the Khibinskiy and Lovozero plutonic massifs contain hydrocarbon gases, close to those of oil-gas areas. These gases contain 80 to 98 percent methane and 20 to 22 percent heavy C_2—C_4 hydrocarbons.

2. Basic to ultrabasic rocks of the Monchegory massif carry no hydrocarbon gases, with the exception of a small amount of methane. The similarity of the geologic structure of the Monchegory and Lovozero massifs suggests that the difference in the hydrocarbon-gas content of their component rocks is the result of a difference in the magma composition.

3. The direct relationship between the chemical composition of rocks and their hydrocarbon gas content suggests an inorganic origin for the latter. The observed regularity in change of the hydrocarbon gas content in the Khibinskiy rocks, as a function of their aluminum content, makes it possible to formulate a fairly well-substantiated hypothesis as to their origin.

An alkaline molten solution had all the prerequisites for generating hydrocarbon gases: a reducing environment and a high temperature. Hydrogen could have been generated by action of water steam on ferrous-iron compounds. This is known to produce hematite or magnetite and free hydrogen. This reaction is reversible and proceeds with a fair intensity up to 500°C.

Methane could have been formed through synthesis of carbon and hydrogen. This is suggested by the relation of the hydrocarbon gas content and the aluminum concentration in rocks. Aluminum is the catalytic agent in this case. Hydrogen and hydrocarbon gases were generated during the intrusion and crystallization of the alkaline molten solution.

4. Alkaline rocks of the Khibinskiy and Lovozero massifs contain reduced bitumens in microfractures and rock pores.

The consistent relation between hydrocarbon gases and bitumen in rock, suggests the latter's formation through polymerization of saturated hydrocarbons.

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Received May 5, 1958

THE PROBLEM OF MIOCENE (PRE-PANNONIAN) VOLCANISM OF TRANS-CARPATHIA [1]

by

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This paper gives a brief review of Miocene volcanism in the Trans-Carpathian foredeep. An attempt is made to establish a general relation between it and the larger province of the inner Carpathians. This relationship enables the author to suggest a stratigraphic position for some of the more important and controversial (as to their age) volcanic beds of the Trans-Carpathia. A precise determination of the age of terrigenous sedimentary deposits which enclose them is not always possible, because of the paucity of faunal data.

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INTRODUCTORY REMARKS

The Carpathian geosynclinal province exhibits impressive evidence of Tertiary volcanism which followed the main tectonic phase of folding. In other words, it is undoubtedly upper Paleogene. Our knowledge of it is far from adequate, especially in comparison with that of Miocene volcanism, even as to the earliest stages of the latter. This is the result of the small distribution of volcanic products among Paleogene deposits and because their study has been systematically made for only a short time.

In the pre-1948 publications, this early stage of volcanism is mentioned by M. Kuthan [38]. He believes that the first phase, in the subject area, opened with basic extrusions of porphyritic and diabase types in the Upper Cretaceous, as indeed was the case. It continued in the Oligocene-lower Miocene with the extrusion of acid rocks such as rhyolites, rhyodacites, and their tuffs. Here we take issue with M. Kuthan's correlation. According to data on hand, the upper boundary of his first phase for the Trans-Carpathian downwarp should be moved considerably higher up, to be drawn somewhere near the Helvetian-Tortonian boundary.

Oligocene tuffs from the Soviet Cis-Carpathian menilite series (Chechva liparite tuffs) were previously described by O. S. Vyalov et al. [5]. If we agree with M. Kuthan [38] that the contemporaneous volcanism was localized in the area of the present Matra Mountains, Hungary, our Chechva tuffs will have to be associated with this stage of volcanism, thus postulating an aerial transportation of ash load over a distance as much as 350 km [20], which is quite possible.

A major phase of volcanism occurred in the Pannonian and persisted to the Pliocene. Its earliest stages, however, seem to have appeared in the lower Sarmatian. At that time, a chain of volcanic ranges was formed, from the Tokay-Preshovo plateau and Vygorlat-Gutin ridge to the Kaliman and Khargite massifs of Romania. In so far as the contemporaneous volcanoes issued mostly products of basic magmas (andesites, basalts, and very rarely acid products, these products naturally could not disperse like the acid pyroclasts and settle over a vast territory. Thus, the volcanism of this last phase differed from the Miocene phase in its lateral extent, but even more so, in time. It is not impossible, however, that evidence of older volcanism is present in the area of younger ranges.

Younger volcanism of the Trans-Carpathia has been already described in adequate detail [20]. Accordingly, we shall only briefly review certain phases of Miocene volcanism, up to and including the Sarmatian.

Miocene tuffs from the Trans-Carpathian Solotvino depression are first mentioned in the 1932-1934 Czech publications (such as D.N. Andrusov's) on coal deposits. In 1937, V. Cechovic studied the lower part of that section. He noted pale-green rhyolite tuffs on top of red, cemented, course, basal

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[1] K voprosu o miotsenovom (dopannonskom) vulkanizme zakarpat'ya.

Miocene conglomerates These tuffs occur locally along the edges of the area and are believed to be correlative with the Helvetian Dej tuffs of the Transylvanian depression.

In 1939-1944, the Miocene cross sections of the Solotvino depression were studied by the Hungarian geologist F. Sentej, who believed that green dacite tuffs were the oldest Miocene rocks of that depression. He assigned a Helvetian age to most of them, with a possible Burdigalian for the underlying rocks.

Systematic studies of the last 11 years resulted in a more complete geologic picture of the Trans-Carpathian depression and in a general stratigraphic section of its Miocene formations. We shall mention here the 1948-1949 tabulations of N.P. Yermakov [12], I.A. Korobkov, and I.B. Pleshakov [13], and I.B. Pleshakov [16]. The last two works became a basis for further study which brought about a partial modification. Papers by V.I. Slavin and N S. Filimonova, 1953 [18]; the 1956 papers by O.S. Vyalov [7], K. Ya. Gurevich [10], N.V. Dabagyan, et al [11], G.N. Grishkevich [8], and I.V. Venglinskiy [4] dealt to various extents with Miocene stratigraphy of the Trans-Carpathia.

The change in ideas regarding geologic structure and relationship of Miocene deposits of the Trans-Carpathian downwarp is of great interest inasmuch as it involves numerous tuffaceous layers which reflect the volcanic processes of the inner Carpathians, from lower Miocene to and including the upper Sarmatian. In addition to the known and published works on this subject, there are new data obtained from field work and especially from drilling.

Without going into details of each of the above-named stratigraphic sections, we shall generalize some of their features. Although we do not know as yet the location of volcanic centers which yielded locally immense thicknesses of pyroclastic material, nearly all geologists working in the Carpathians are inclined to place these volcanoes in the Trans-Carpathian interior downwarp. Their explosive force was so great (of the Katmai type) that, despite the settling of most of the explosive products in their vicinity, a considerable portion might have been transported as far as 150 to 250 km. Consequently, tuffaceous layers are known not only from the inner Carpathian province (Czechoslovakia, Hungary, Soviet Trans-Carpathia, and Transylvania) but also from the entire Cis-Carpathian foredeep, from Romania to Poland, and even in the west of the Russian platform, some distance from its edge.

Considering that evidence of volcanism is very consistent throughout the section, it is very tempting to tie it to definite stages of volcanic activity. The genetic kinship of various volcanic layers is confirmed, to a considerable extent, by detailed petrographic study. This, in turn, may facilitate the solution of a converse problem of using the volcanic stages so established for a vast area, as a means of correlating the local stratigraphic columns, tentatively established for each of the above regions. In other words, such a study may be helpful in compiling a single correlation, be it only for the Trans-Carpathian and Cis-Carpathian foredeeps. Such a composite stratigraphic column is still lacking and local correlations need much refining.

Study of volcanism in the Soviet Carpathians is being carred out on a large scale, but is far from complete. However, the data on hand make it possible to make certain statements on the distribution of pyroclastic rocks throughout a relatively large area, mainly on the position of tuffaceous layers among Miocene deposits of the Trans-Carpathia.

POSITION OF VOLCANIC FORMATIONS IN MIOCENE DEPOSITS OF TRANS-CARPATHIA

As early as in 1948, N.P, Yermakov assigned the Mt. Klobuk (area of the village of Zadneye) tuffs to the Aquitanian-Burdigalian, the oldest Miocene of the Trans-Carpathia. The basis for this was a Tertiary fauna found there by V.I. Slavin and identified as Aquitanian by T.N. Baykovskaya. In 1952, a stratigraphic hole, Danilovo No. 1, was drilled in the Solotvino depression. At 1203-1916.7 m, it penetrated a nearly solid sequence of tuffs and tuffites with some, although fairly thick, mudstones, siltstones, and sandstones. K. Ya. Gurevich [10] identified this sequence as the so-called Danilovo formation. The tuffs are nowhere exposed and, naturally, are not shown in the stratigraphic columns of I.B. Pleshakov. Incidentally, Aquitanian tuffs, designated for the Trans-Carpathia by N.P. Yermakov, are not shown there, either. In 1953, V.I. Slavin and N.S. Filimonova [18] noted that the Miocene section of the northeastern part of the Solotvino depression opens with an Aquitanian-Burdigalian arenaceous-argillaceous formation. It is followed by lower dacite tuffs (Danilovo formation). According to those authors, it is exposed in the Borshava valley (by the Klobuk Mountain) and on the southeastern edge of the depression, in the Apshitsa basin. This formation, which they named, Klobuk, carries an assemblage of warm-water Miocene fauna. As a result, V.I. Slavin and N.S. Milimonova, like N.P. Yermakov, assigned the Klobuk tuffs to the lower Miocene.

K. Ya. Gurevich [10] assigned the Danilova formation to upper Burdigalian, after L.S. Pishvanova and N.V. Dabagyan had identified its faunal assemblage. A contemporaneous paper by N.V. Dabagyan [11] assigned this formation to the Burdigalian-Helvetian. O.S. Vyalov [7] puts it in the Aquitanian-Burdigalian sequence, above the tuffs, together with the Tereblino series. No one, except for V.I. Slavin and N.S. Filimonova, associates Kobluk tuffs with the Danilova, with the latter assigned a different age by various authors.

Various authors are also at odds as to the presence of lower Miocene sediments not only in the Soviet Trans-Carpathian but throughout the Pannonian basin as well. This is especially true for the Aquitanian. Thus, according to L. Straus [51], "the very separation of the Aquitanian is questionable" for the Pannonian basin, let alone the division of lower Miocene into the Aquitanian and Burdigalian. He begins his own stratigraphic section with the Burdigalian. To be sure, there is evidence of an Aquitanian regression in the area of Byukk and especially of Shalgotar'yan. I. Csepreghy-Mezneries [30] and E. Szotz [52] substantiate the presence of the Aquitanian in other parts of Hungary by a marine fauna, mainly Glycymeris obovata. E. Vadasz [55], E. Lengyel [39], and the above-named authors place Burdigalian deposits with "lower rhyolite tuffs" developed in the areas of Dunazug, Chergat, Matra, Byukk, Shayo River, and Bodva River, etc., i.e. mostly along the western border of the Pannonian basin.

The distribution of lower Miocene deposits throughout eastern Slovakia [46], too, certainly does not justify the earlier concept that the Neogene province "subsided more or less evenly in the beginning of the Neogene and was gradually filled with Miocene sediments." The main role in the distribution of sediments was played by long and narrow belts along faults which originated in the lower Miocene. Aquitanian deposits have not been definitely identified within eastern Slovakia.[1] Here, the lower Miocene includes, without much justification, Burdigalian-Helvetian formations which carry montmorillonite beds near the villages of Fintitse and Kapushany. M. Kuthan [38] assigns even extrusive rhyolites to the Burdigalian. However, the sandstone fauns which enclose the montmorillonites is poorly preserved and does not definitely bespeak their Burdigalian age. V. Cechovic [28] correlates these rocks with rhyolite tuffs of the Solotvino depression, which he had studied before [27]. He assigns both of them to the Helvetian.

Thus, Burdigalian volcanism is established more or less definitely only for the island mountains of Hungary and in an area northeast of the Danube. Its intensity, to judge from the thickness of pyroclastic products (about 30 m; [55]), was not great.

An unequivocal determination of the age of the Danilovo tuffs is much more difficult. It is difficult to imagine that a very intensive and obviously prolonged inner-Carpathian volcanism, which left behind 700 meters of nearly entirely pyroclastic material, did not extend over a wider area including Romania, Czechoslovakia, and Cis-Carpathia. It is significant in this connection that the well-traceable tuffs, 2 to 5 m and rarely as much as 25 m thick, occur in terrigenous deposits throughout the Soviet Trans-Carpathia, from the Helvetian up. Assuming that they were transported there through the air, even with explosions somewhat decreasing in intensity with time, it is very strange that lower Miocene tuffs correlative with the Danilovo are absent, for instance, in the Vorotyshcheno series of the Cis-Carpathia. The authenticity of tuffs identified by Yu. Tokarskiy [54] from the Vorotyshcheno series in the Stryy area requires verification. At the same time, two layers of acid tuffs, with the upper one no less than 25 m thick and the lower one apparently somewhat thinner, were identified in the overlying Helvetian Stebnik formation (series) (village of Krasnoye in the Nadvornaya area). These tuffs were first noticed by M. Kamienskiy [37] and described later by D.P. Bobrovnik [1], V.S. Sobolev, et al [19, 20], and S.M. Korenevskiy [14].

A fairly thick (from 70 to 150 m) familiar Dej tuff layer is known from Romania. However, inasmuch as this, the lowest, tuffaceous layer is spread over a vast area, virtually covering all of the Transylvanian depression and the Apusheni massif, it is not assigned the same stratigraphic position everywhere. Thus, in the Transylvanian basin, the Dej tuffs most commonly overlie the Corod and Hidal'mash beds, thought to be Burdigalian. A Helvetian age is therefore assigned to the Dej tuffs. However, I. Voitesti [56] and R. Ciocardel [24, 25] place it at the Burdigalian-Helvetian boundary. I. Atanasiu [23], the author of a comprehensive review of Neogene volcanism of Romania, assigns to the Helvetian the Zlatnoy (Apusheni massif) rhyodacite extrusives and the Transylvanian basin Dej tuffs.

A "middle rhyolite tuff" layer is separated in the Helvetian of Hungary, in the Cherget, Matra, and Byukk massifs, and at the Shayo River [30, 52, 53]. In addition, there are

[1] A minor Aquitanian transgression was previously established for southern Slovakia [44]. In a later work by J. Senes [47], the Burdigalian includes all of the Aquitanian, in the former meaning of the term, with the alternation of rhyolite tuffs placed in the upper Burdigalian.

Helvetian flows of rhyolites at Shekshard (Mechek) and dacites (Dunazug and Borzheni), as well as their tuffs [39, 55].

In eastern Slovakia, according to V. Čechovic [28], "the first extrusions of rhyolites and the rhyodacite tuffs occur approximately at the Burdigalian-Helvetian boundary and are concentrated in Helvetian deposits." The same idea is stated in his other paper [29].

From these correlations, we are inclined to agree with I.B. Pleshakov that the volcanism responsible for the Danilovo tuffs developed mostly in the Helvetian, although possibly initiated in upper Burdigalian. It appears that only those major volcanic eruptions responsible for the accumulation of a 700-meter thick sequence of the Danilovo tuffs could have been the source of ash some 120 km away and of fairly thick (35 to 40 m) Cis-Carpathian tuffs.

The next fairly intensive phase of volcanism is identified in the Solotvino formation and is approximately dated as middle and even upper Helvetian [10]. That formation is now correlated with the Cis-Carpathian Stebnik formation, with a Helvetian age for both, with a fair degree of certainty [7, 11]. The Solotvino tuff layers attain 40 to 50 m with an overall thickness of about 80 m. The tuff beds are so consistent laterally that I.B. Pleshakov was able to use them as markers for his subformations. They are exposed in the vicinity of Solotvino and are penetrated in the Danilovo No. 1 borehole.

The best known is the so-called "Novoselitsy dacite tuffs" layer. It has a definite position in the Solotvino-depression Miocene, where it is a good marker. Its average thickness is about 200 m, varying considerably from place to place, usually in the direction of thinning. These tuffs crop out in two bands: one is traced to the south, along the salt-diapir folds; the other to the north, along the line dividing Miocene deposits of the depression and Cretaceous-Paleogene flysch of the Carpathians with outcrops of Jurassic limestones.

The position of the Novoselitsy tuffs in the stratigraphic column of the Trans-Caucasian downwarp is still controversial. I.B. Pleshakov [13, 16] places it at the base of the Khustets formation whose age he gives as upper Helvetian. N.V. Dabagyan et al [11] point out that "a correlation of the Khustets-microfaunal assemblage with that of Amusium denudatum beds (i.e., with the Cis-Carpathia, V.P.K.) suggests its lower Tortonian age." In determining the position of the Khustets formation, K. Ya. Gurevich [10] also refers to the above-cited authors. O.S. Vyalov [7] writes that "the question of the age of the Khustets formation must be re-examined," and that the roster of its microfauna "suggests rather. . . . their lower Tortonian age." In a revised section, he places the Khustets formation (raising it to the rank of series) in the Tortonian.

In concurring with this opinion, we again turn to the analysis of contemporaneous volcanics. If we leave, with I.B. Pleshakov, the Novoselitsy tuffs in the Helvetian, a Tortonian volcanism in the Solotvino depression would have to be virtually ruled out, since his stratigraphic columns contain but a single two-meter thick tuff layer in the Tyachevo formation. It is true that I.B. Pleshakov [16] noted as early as 1949 that a 10 to 12-meter-thick tuff layer, exposed in the Klobuchek Mountain on the left bank of Borshava, appeared to be correlative with tuffs exposed in the upper Osava basin. This tuff layer carries a pelecypod fauna which puts it in the lower Tortonian (these are the tuffs which other students assign to lower Miocene; [12, 18]). Until very recently, nothing was known of Tortonian tuffs in the northeast of the Mukachevo trough. Only in the Beregovoye Kholmogor'ye (river highlands) area were Tortonian andesites identified, not as intrusive sills but as lava flows, because of the associated andesite tuffs.

More detailed study in the Vyshkovo area and north of there (village of Bushtino) just recently, revealed some 400 m of sedimentary rocks below the so-called Fenesh formation of I.B. Pleshakov. The lower part of the sedimentary sequence carries isolated tuff layers, with a thick -- 100 to 170 m -- layer of acid tuffs at its base. The latter includes a 20 to 30 m thick arenaceous-argillaceous bed. Its microfaunal assemblage, identified by I.V. Venglinskiy [3], links this layer with the Teresven series, i.e., with upper Tortonian.

The 1954-1956 drilling in the Mukachevo trough revealed the presence of acid tuffs in upper Tortonian of many localities. Thus, borehole No. 6 (Kamenka) penetrated five tuffaceous layers, in the interval of 610 to 722 m. These layers were 2 to 7 m thick (total thickness about 25 m) with thicker layers gravitating toward the lower part of the interval. In the fall of 1956, borehole No. 24 (Irshavy area) penetrated at 940 to 955 m a massive layer of white liparitic tuffs in upper Tortonian sediments.

Turning now to the adjacent Soviet Trans-Carpathia, we see that acid volcanics (chiefly acid tuffs with subordinate extrusives of a dacite and liparite type) are common in the Tortonian of eastern Slovakia. Thus, rhyodacites are observed in the Tortonian, north of Vranovo. According to J. Senes and others, there are extrusive equivalents of tuffs and

tuffites exposed at Nizhniy Grabovets (10 to 14 m thick), Vranovo, Kuchino, Zlatnik, and elsewhere.

In the same area, B. Lesko [40] identifies lower Tortonian rhyodacites and their tuffs, 20 to 50 m thick, and upper Tortonian "younger rhyodacites." In the eastern-Preshovo-Tokay foothills and in the northwestern-Tisza plain, the tuff layer is as much as eight m thick. Acid tuffs occur northeast of the Zemlinskiy island. F. Cech [26] notes that lowermost (marine) Tortonian deposits in the area of Banovets, Mikhaylovtsy, and Strazhskoye, include two volcanic layers, the lower one of which is made of orthoclase-plagioclase rhyolite, and the upper of rhyodacite tuff. J. Slavik [48] tells of the finding of an acid-tuff layer in Tortonian deposits near Zhidlokhovits, Moravia, i.e., much farther west of the eastern Slovakia area of these volcanics. In the northern Danubian plain, L. Ivan [36] conditionally assigns to the Tortonian the so-called "underlying andesite," with a reservation that the latter may be as young as lower Sarmatian. Tortonian andesite tuffs at the villages of Vyshniy and Nizhniy Chay are described by J. Svagrovsky [50]; and those between Gerlyany and Bidevtse by J. Gasparik [31]. The possibility of andesite-magma extrusions for the Tortonian is suggested by J. Senes [45]. However, he is inclined to assign them to the lower Sarmatian. The appearance in the heavy fractions of Tortonian sediments from the Vagranovets, Zamutovka, and Koshitse formations, of such minerals as monoclinic amphiboles and rhombic pyroxenes (possible products of andesite disintegration) also points to andesite extrusions.

The Slovakian geologists, after M. Kuthan, put much emphasis on Tortonian andesite extrusions which are supposed to have occurred in many areas (andesite phase I; I. Svagrovsky [49]). However, J. Senes [44] states that "the lack of paleontologic data has precluded an exact age determination (Tortonian-Sarmatian-Pannonian) for certain andesites, rhyolites, and dacites, chiefly from eastern Slovakia."

Tortonian rhyolite tuffs are developed nearly everywhere in the chain of island mountains of Hungary, from Mechek to Tokay. Data on them are contained in many publications, of which a general summary by E. Vadasz [55] is the most complete. It is of interest that below acid tuffs in the same areas (with the exception of Bakoni), there lies a formation of pyroxene, less commonly of amphibole, andesites, and their pyroclasts.

The Transylvanian-Basin Tortonian includes, in the unanimous opinion of geologists, the Gadareni acid tuffs usually overlain by the Girish-tuffaceous layer. For the Apusheni region, C. Gheorghiu [32] designates as Tortonian, the Deva-area dacite tuffs (Muresh River); I. Atanasiu [23] assigns to the Tortonian the Draitsa dacites and underlying flows of the Roshia andesites. M. Ilie [35] describes Tortonian dacite tuffs from the Rakosh area (the Oltul River valley), 100 to 150 m thick, and tuffs from other localities [33, 34], which he correlates with the Dej tuffs. I. Patrut [43] points out the similarity between tuffs from Dej and the Beklyany area (40 to 45 m thick), and places them at the Helvetian-Tortonian boundary. It appears that the age of the Dej tuffs is not quite clear. On the basis of published data on Romania, it may be surmised that these authors either are wrong in associating tuffs from the above-named localities with the Dej tuffs, or else they assume the latter to be older than they really are.

Tortonian tuffs are known from many localities of the Soviet Cis-Carpathia, especially in the Outer Belt and on the platform. Data on them, mostly scattered, are contained in both the Polish (M. Kamiensky [37]), and Soviet literature (D.V. Gurzhiy et al [9]; S.M. Korenevskiy [14]; M.B. Ripun [17]; L.G. Tkachuk et al [21]). According to these authors, tuffs occur both in subanhydrite Baranovo beds of the Balich series (Uger series of O S. Vyalov [6]), assigned to lower Tortonian, and in the upper-Tortonian Kosvo (Pokuty) formation. They also occur in the Konsk beds (more probably Konsk-Buglov) transitional from the Tortonian to Sarmatian. We have collected Tortonian tuffs from well cuttings and cores throughout a comparatively large area of the Russian platform, locally at a considerable distance from its southeastern edge.

The area of distribution of Tortonian tuffs apparently is very large, extending approximately from Chernovtsy in the southeast to Tarnow in the northwest. W. Parahoniak [42] studied a number of exposures of tuffs and tuffites from the Tarnow and Bokhnya areas. They all occur in the Khodenitsy beds whose age is determined as lower Tortonian. The tuffaceous layer here is 3 m thick on the average. In addition, the same author describes tuffs from the Pil'zne and Mil'tse areas, occurring in the Krakow clays of upper Tortonian age. In their chemical composition, both belong to acid tuffs similar to our Cis-Carpathian tuffs. S. Aleksandrowizc et al [22] encountered similar tuffs in the Khodenitsy beds near Pinchovo.

Coming back once more to the position of the Novoselitsk layer, it should be noted that until a very recent time there was a discrepancy in out thinking on the Tortonian volcanic phase in Soviet Trans-Carpathia. We have

seen that evidence of this phase, in the fairly thick tuffaceous layers and, in places, even in their extrusive equivalents such as rhyolites and dacites, occurs literally everywhere throughout the adjacent regions, including the Cis-Carpathia. Yet, for some reason, they were omitted in the Miocene stratigraphic sections for the Trans-Carpathia. The Tortonian-volcanic stage undoubtedly was highly intensive, and its traces, before developing in the Cis-Carpathia, should have been preserved in the Solotvino depression, even if the volcanic centers had been located somewhere in the adjacent regions. In this respect, too, we fully concur with the above-named authors [7, 11] who assign the Khustets formation (series), along with the Novoselitsy-tuffaceous layer, to the lower Tortonian.

Products of the Sarmatian volcanism are preserved in many localities of the Trans-Carpathian downwarp and are observed in the Solotvino and Mukachevo troughs. They are mentioned in early stratigraphic sections [12, 13]; microfauna of the enclosing rocks was studied by I.V. Venglinskiy [2]. The 50 to 60-m thick "Capucine" layer of N.P. Yermakov, which is a lower tuffaceous layer of the Vyshkovo area, is placed by I.B. Pleshkov at the base of his Bogosh formation. The upper layer (15 to 30 m thick) he places at the base of the Techen formation. These two layers are separated by approximately 150 m of sedimentary terrigenous rocks. Sarmatian deposits of the Lipshany area also contain several isolated, relatively thin layers of acid tuffs which are usually correlated with the Vyshkovo tuffs.

The Beregovskoye Kholmogor'ye section includes, above the previously mentioned Tortonian andesites and their tuffs, the so-called lower sedimentary sequence, assigned to the Tortonian by I.V. Venglinskiy [4]. It is overlain by a lower layer of liparitic tuffs, varying laterally from 25 to 250 m in thickness. It is Sarmatian in age, according to some; and considered probably Tortonian by some others. The finding of an Ervilia trigonula Sok. fauna in these tuffs suggests, as noted by G.N. Grishkevich ([8], page 176) that they are younger than the lower Tortonian and "most probably, judging from the age of overlying beds, this is the base of the Apshin series -- the Baskhev and Beshikur formations," i.e. upper Tortonian. The lower tuffs are followed, above the sedimentary sequence, by other liparitic tuffs of the upper layer, as much as 100 m thick. Both these sequences are lower Sarmatian (Abro beds of G.N. Grishkevich [8]).

Recent drilling on the northeastern edge of the Mukachevo trough and in the Irshava hollow penetrated a number of tuffaceous layers, from 1 to 4 m thick on the average, in places as much as 10 m. Most belong to the Zalugi beds (lowermost Sarmatian) and in part to the Lukov beds (upper part of the lower Sarmatian). The northernmost known Sarmatian tuffs are encountered a little south of Mukachevo, at a depth of 345 to 365 m.

Three tuffaceous layers are identifiable in the Lukovo beds of the Irshava hollow: first, as much as 3 m thick, at 215 to 240 m; second, as much as 6 m thick, at 470 to 494 m; and third, 3.5 m thick, at 544.4 to 578.4 m. One borehole penetrated three tuffaceous layers in the Zalugi beds, at 520 to 580 m. The two upper ones are 2 to 2.5 m thick, each; the lower one, about 25 m thick, is made up chiefly of tuffaceous material with subordinate clays in thin layers. It is not impossible, although subject to further study, that liparite-dacite tuffs and their intrusive equivalents, observed at the base of the Vygorlat-Gutin volcanic-range section (Pannonian) in the Perechino area, also may be contemporaneous with the Beregovskoye Kholmogor'ye volcanics. Unfortunately, they there rest directly upon Cretaceous-Paleogene flysch, and their relationship with Miocene sediments have not been as yet established.

Thus, the presence of acid products of volcanism in Sarmatian sediments has been definitely established for the Solotvino and Mukachevo troughs. The magnitude of this volcanism, as far as the thickness of their pyroclastics and the area of their distribution is concerned (of which more below), was quite impressive.

Less is known of the age of andesites with coarse hornblende crystals, as much as several centimeters long, which appear in the area of Lipsha and Zadneye. Among Sarmatian deposits of the Mokanu River area, there are two fairly thick andesite beds; the lower, 35 to 40 m thick; the upper, 80 m. They are separated by 55 to 60 m of clays, sands, and friable sandstones. Petrographically, these andesites are identical with those making up a number of domes such as the Klobuk Mountain, Malyy Klobuk, Klobuchek, etc. The author studied intrusive contacts of these andesites with the surrounding rocks, for example along the Bystrik River, between Zadneye and Lukovo. The intrusive position of andesites is confirmed also by numerous small xenoliths of hornstone formed by the metamorphism of the enclosing rocks. No substantial alteration of argillaceous rocks at the contact with definitely intrusive andesites has been found, however, except for some consolidation in the contact zone. No xenoliths have been found in bedded andesites along the Mokanu. The near-contact alteration of enclosing rocks was also lacking. Considering the fact that sediments which enclose these comparatively thick andesites were

subjected neither to thermal nor mechanical action, and that they are rather poorly consolidated, it seems more reasonable to assume that, along with domes and dikes in flysch, andesites may have occurred in flows into a water basin such as the Mokanu. In that case, the age of all these andesites of a similar composition may be unequivocally established as Sarmatian, more precisely, the upper lower Sarmatian.

A sarmatian andesite volcanism in the Trans-Carpathia is, then, conjectural for the time being, although very probable inasmuch as it undoubtedly took place in the neighboring eastern Slovakia. A 1955-geologic map of eastern Slovakia, by B. Lesko, J. Senes, and J. Svagrovsky, shows Sarmatian andesites (called by them, autometamorphic) and associated pyroclastics in the Vinny area; an extended massif is located south of Zlata Banya; and some of their outcrops is located north of Preshov. L. Ivan [36] assigns to upper Tortonian the andesites of the northern Danubian plain, with the reservation that they may belong at the base of the Sarmatian, because of the overlying tuffites with a Sarmatian fauna. J. Senes [45] mentions that most andesite flows are associated with the Sarmatian, with only insignificant Tortonian andesite eruptions whose flows are now concealed by younger volcanics of the Preshov-Tokay highlands.

Lower Sarmatian rhyolites and their pyroclasts are known from many eastern Slovakian localities. The same map shows them as a chain of small massifs stretching along the west slopes of the Preshov highlands, from the above-named Sarmatian andesite massif at Zlata Banya to the south boundary and farther on to Tokay. A number of small rhyolite massifs occurs also east of there, in the Mikhlovtsy area (valley of the Laborets river).

The upper Sarmatian sequence of eastern Slovakia, about 300 m thick, formed in a fresh-water basin, is also rich in volcanic material, chiefly tuffaceous (Sechovtse area, west of Koshitsa, in the valleys of the Gornad, Olshava Rivers, etc). Nevertheless, the separate upper-Sarmatian volcanic phase of Czech geologists should be viewed with caution. As we understand it, this phase should probably be assigned to the Pannonian.

A correlation of data from a larger province reveals that acid extrusives and explosions of the second phase came to an end, mostly in the lower Sarmatian when the third andesite phase of volcanism began to develop.

The Sarmatian-volcanic activity in the Semigrad Mountains and Transylvanian basin is marked by a consistent layer of the Bazna dacite tuffs which are mentioned by many students. In the Apusheni massif, C. Gheorghiu [32] assigns to the Sarmatian the andesite "pyroclasts of the Gurazada type," whereas M. Ilie [33] places there, the Sarmashel tuffs whose age is determined as transitional from the Sarmatian to the Pannonian. For the Transylvanian basin, I. Atanasiu [23] cites andesite tuffs, locally with amphibole, in the Mt. Blaj area; their extrusive equivalents are probably andesites at Breaz and Zhiva.

Rhyolitic rocks of Sarmatian times are fairly well developed in the region of the island mountains of Hungary: Chergat, Matra, Byukk, Tokay, and others. At the same time, this area witnessed the appearance of the third, andesite, phase which is well developed in the Pannonian and Pliocene.

Several fairly consistent lower Sarmatian acid tuff layers, 1 to 4 m thick, are well traceable in the northwestern part of the Cis-Carpathian downwarp inner zone and on the adjacent platform. There are no descriptions of them in the literature, unless it be G. A. Nechayev's description of upper Tertiary (?) tuffs [15] which may be Sarmatian, like other tuff outcrops in the Razvadovo area. The lower-tuff layer belongs to the Konksk (most kilely Konksk-Buglov) deposits; the upper, to the Volyn. They are separated by a 50 to 90 m thickness of arenaceous-argillaceous rocks.

The data cited in this brief review are summarized in Table 1. They do not pretend to embrace the fullness of Miocene (pre-Pannonian) volcanism in the Trans-Carpathia, since our knowledge of it increases all the time. The correlation of individual layers is somewhat conditional, in a number of instances; first, because of the difficulty of dating even lithologically similar units containing an inadequate fauna; second, because of the difficulties in the tying together of stratigraphic Miocene systems over the immense distances of the outer and inner Carpathians. Only the most important volcanic layers have been used, more or less definitely correlated by various authors.

Localities of these rocks, chiefly in the inner Carpathians, are given on the index map (Fig. 1) which also shows those extrusives found below the surface. The distribution of younger (Pannonian) volcanics for the same region is shown in Figure 2, borrowed from M. Kuthan [38].

SUMMARY

The Carpathian orogeny, initiated in the upper Paleogene, undoubtedly was accompanied

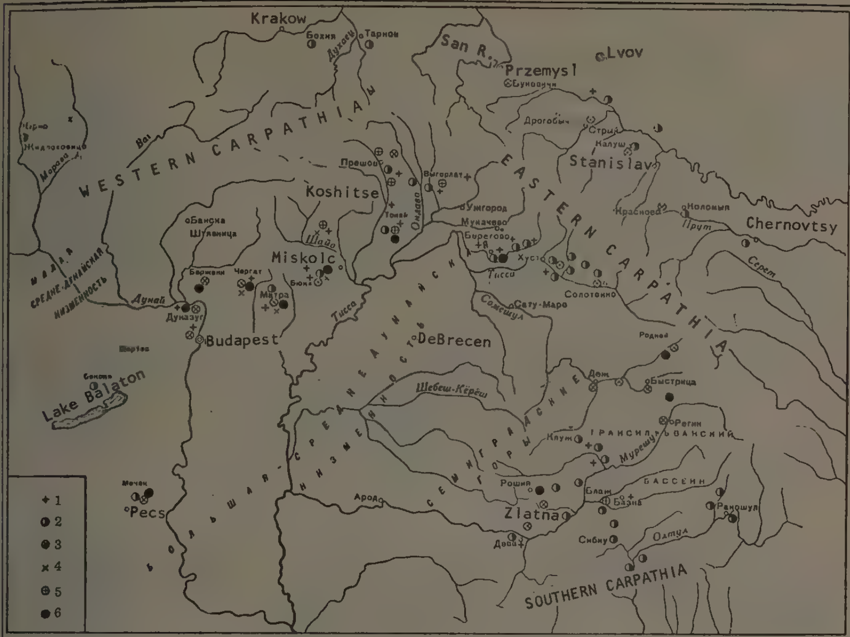

FIGURE 1. Distribution of Miocene volcanics in the Carpathians.

Rhyolite and dacite tuffs, their less common flows and extrusive equivalents: 1 -- Sarmatian; 2 -- Tortonian; 3 -- Helvetian; 4 -- Burdigalian; Andesites and their tuffs; 5 -- Sarmatian; 6 -- Tortonian.

FIGURE 2. Distribution of upper Miocene volcanics in the inner Carpathians, after M. Kuthan.

Table 1

Correlation Sections For Volcanic Rocks of the Carpathian Province

Stage	Horizon	Solotvino Trough	Mukachevo Trough	Romanian Trans-Carpathian
Lower Sarmatian	Second	Hornblende andesites of the Klobuks and Mokanu River(?) (40 to 75 m). Techen rhyolite tuff (15 to 30 m).	Rhyolite tuffs of the Lukovo beds (3 to 5 m).	Andesite tuffs of the Sarmashel (?), Gurazada, and Blaj areas. The Bazna dacite tuffs.
	First	The Bogosha (Capucine) rhyolite tuff (50 to 60 m).	Rhyolite tuffs of the Zaluga' beds (as much as 25 m). Rhyolites and their tuffs, Beregovo (more than 100 m).	Rhyolite tuffs of the Oltul valley and Barsau area.
Tortonian	Second	Lower (third) rhyolite tuff layer of Vyshkovo area (100 to 170 m).	Beregovo rhyolite tuffs (lower, as much as 200 m). Tuffs of the Mukachevo trough and Irshava hollow (15 to 25 m).	Dacite tuffs at Girish and Gadareni. Dacites at Draits.
	First	Novoselitsy plagiorhyolite tuffs (as much as 200 m) and "Klobuk" tuffs (10 to 12 m).	Andesite and their tuffs of the Beregovo area (?) (more than 300 m).	Older andesites at Roshia. Dacite tuffs of the Deva and Rakosha areas (as much as 150 m).
Helvetian	Second	Rhyolite tuffs of the Solotvino formation (total as much as 80 m).		Rhyodacite at Zlatna. Dacite tuffs at Beklyany.
	First	Most of the rhyolite-tuff explosions of the Danilovo formation (more than 700 m), initiated probably in the Burdigalian.		Dacite tuffs at Dej (70 to 80 m).
Burdigalian				

Table 1, continued

Correlation Sections For Volcanic Rocks of the Carpathian Province

<table>
<tr><th>Hungary</th><th>Eastern Slovokia</th><th>Soviet Cis-Carpathia</th><th>Polish Cis-Carpathia</th></tr>
<tr><td>Rhyolite and their tuffs in the area of the island mountains Dunazug, Chergat, Matra, Byukk, and the Shayo River.</td><td>Products of andesite and rhyolite extrusions of the Koshitse hollow, area of Sechovtse, Gornad and Olshava Rivers (upper tuffites).</td><td>Upper rhyolite-tuff later of the Volyn layer (0.3 to 0.7 m).</td><td></td></tr>
<tr><td>Rhyolites (normal and perlitic) of the Tokay mountains.</td><td>Rhyolite tuffs and rhyolites of the Koshitse hollow, Mikhaylovtsy, Preshov highlands, etc (lower). Andesites and their tuffs at Mikhalovtsy, Zlata Banya, Preshov, and other areas</td><td>Rhyolite tuffs of the outer downwarp zone (1.5 to 2 m). Konsk (Concha?) (Konksk-Buglav) tuffs.</td><td></td></tr>
<tr><td rowspan="2">Rhyolite tuffs of the island mountains Mechek-Tokay (near Chergat and Dunazug). Pyroxene (less commonly amphibole) andesites and their tuffs in the same area. Norgrad dacites and Matra-rhyolite tuffs.</td><td>Rhyodacite tuffs of the Kuchino, Straj, and Mernik areas. Andesites and their tuffs of the Preshov Mountains, Modry Kamen', and others.</td><td>Rhyolite tuffs of the Chernovtsy and Stanislav provinces (1 to 4.5 m). Tuffs of the Kosovo (Pokuty) formation (0.75 to 2 m).</td><td>Rhyolite tuffs from the Pilzne and Miltse area.</td></tr>
<tr><td>Rhyodacites and their tuffs in areas of N. Grabovets, Kuchino, Zlatnik, Vranovo (20 to 25 m). Preshov foothills, at Zemplinsk Island. Tuffs of Zhidlokhgovits, Moravia.</td><td>Rhyolite tuffs of Baranovo beds in the downwarp zone (5 to 6 m) and on the Russian platform.</td><td>Rhyolite tuffs of the Khodenitsy beds in the Tornov and Bokhnya areas (as much as 3 m) and at Pinchovo.</td></tr>
<tr><td rowspan="2">Rhyolites and their tuffs in the Mechek, Buda-Dunazug. "Middle rhyolite tuffs of the island mountains Chergat-Matra-Byukk and Shayo River.</td><td></td><td>Rhyolite tuffs of the areas of Krasnoye (25 m). Bunevichi (6 to 6.5 m), etc.</td><td></td></tr>
<tr><td>Rhyolite flows at Fintise. Montmorillonite beds in the Finitse and Kapushany areas.</td><td>Rhyolite tuffs at Krasnoye (as much as 40 m ?). Tuffs at Ugersko, Bunevich, etc.</td><td></td></tr>
<tr><td>"Lower rhyolite tuffs" (less commonly bentonites) of the same areas.</td><td></td><td></td><td></td></tr>
</table>

by a revival of volcanism, especially marked in the northwest corner of the Pannonian basin. Thus, as early as the Bartonian, relatively thick andesite tuffs are noted [53] among shallow-water organic limestones of the Chergat, Matra, and Byukk massifs. An even higher intensity of andesite extrusions is marked in the same areas and especially in the Byukk at the base of the Ruppelian [53, 55], by tuffs in the "Kischtzel clays." According to E. Cech, the volcanic centers were located on the Lakhotsa (Matra) Mountain and presumably in the Borzheni area. E. Lengyel [39] associates with upper Oligocene time, the extrusions of garnet-biotite and pyroxene-amphibole andesites and their tuffs in the Dunazug area. B. Mauritz [41] also puts the Matra biotite-amphibolite extrusive andesites at the Eocene-Oligocene boundary. Therefore it appears to us that there is no need for lowering the initial volcanic stage down to the Cretaceous, as is done sometimes. The Mesozoic witnessed a quite different, the so-called ophiolite volcanism, developed not only throughout the Carpathians but in a number of other regions, as well.

The main stage of folding responsible for the formation of the central Carpathian structures, and the ensuing sinking of mountain foredeeps in the beginning of the Miocene, could not but be accompanied by a revival, and most probably by initiation, of new magmatic hearths, with further differentiation of magmas in the direction of increasingly acid components.

The volcanism of the first stage, which opened with extrusions of mainly andesitic lavas, gave place to acid explosions. The latter continued into the Helvetian, when they were widely developed and resulted, locally, in large accumulations of tuffs.

Momentous tectonic events took place at the Helvetian-Tortonian boundary (Styrian phase of Stille). The development of the Carpathian foredeeps of that time is well described by O.A. Vyalov [7]. Volcanic activity grew in proportion to the mobility: the inner Carpathians witnessed a new phase of volcanism marked by andesite extrusions in the peripheral parts of the Pannonian and Transylvanian basins. Our own equivalents are the Beregovskoye Kholmogor'ye andesites. Extrusive andesites are followed by acid explosives (less commonly extrusives), terminating in the lower Sarmatian. The third, Pannonian, volcanic phase was not discussed in this paper. On the whole, it was a continuation of the second phase. The fourth phase is found to be incomplete [20].

A correlation of principal tuffaceous layers in the Trans-Carpathian Miocene deposits with those from the inner Carpathians and the Cis-Carpathia suggests that some of the volcanic paroxysms within an individual phase were reflected throughout vast provinces. Study of sections, such as for the Solotvino trough or the Cis-Carpathia, gives an impression that, from upper Paleogene and into lower Sarmatian, only the acid-magma extrusions took place. Indeed, they were in the nature of violent explosions sending ash, chiefly through the air, over long distances, even into the southwestern terminal of the Russian platform. On the other hand, the distribution of the andesite volcanic products was greatly more localized, because of their liquid nature which prevented them from being overstrained at the instant of eruption. Of course, the overall bulk of basic-magma products in the total amount of the inner Carpathian extrusives was appreciably smaller. It is also clear that any ash may have settled over the Carpathians proper, as well, as evidenced by the tuffaceous layers among upper Paleogene deposits,[1] i.e., where they were caught in water basins and were buried there. However, the Carpathians subsequently underwent a continuous uplift and were strongly eroded; as a result, we have not yet found any traces of volcanic material there.

Only general statements may be made on the location of volcanic centers in the inner Carpathians. As already noted, Oligocene volcanism was most likely concentrated in the region of the island mountains of Hungary, northwest of the Danube. It probably persisted there into the Miocene, as shown by lava flows of the same composition as the pyroclasts. Another contemporaneous volcanic center must have existed in the Mt. Mechek area, where Helvetian rhyolites and their tuffs, along with Tortonian andesites and rhyolite tuffs, are observed.

Solely because of the great thickness of the Danilovo tuffs in the Solotvino trough (700 m in a borehole) a Helvetian-stage volcanic center may be postulated either there or not far from there -- perhaps in Romania--inasmuch as these tuffs thin down abruptly toward the periphery of this area. (For the northern part of the Transylvanian basin, the Dej tuffs are estimated to be 70 to 80 m thick; they are much thinner in the Apusheni basin; tuffs of the Cis-Carpathian Stebnik formation are more than 30 m thick at Krasnoye, with 5 to 6 m at Bunevichi, and several

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[1] There are published data on these tuffs and bentonites for the Transylvanian basin (A. Koch), Hungary (W. Ulig, B. Mauritz, E. Vadas, E. Cech, F. Sentes), Soviet and Polish Cis-Carpathia (O.S. Vyalov et al, M. Ksionjkevich and T. Viser). However, we are not concerned here with a detailed study of this subject.

meters thick in eastern Slovakia.) The radius of the ash-distribution area, as determined by known occurrences, was 150 to 180 m on the average, with the maximum of about 300 km.

More definite statements may be made on the Tortonian volcanic centers which supplied the materials for andesite magma. They have been established in the Apusheni massif; in the Beregovskoye Kholmogory'ye, on our side; to a smaller extent in eastern Slovakia, also in the island mountains of Hungary. It is much more difficult to designate the acid-magma sources for these regions. Rhyolite and dacite flows and extrusions are known only from the Apusheni massif (Draitsa dacites) and eastern Slovakia (solely from B. Lesko's data [40], where they rest, of course, near former volcanic centers. Yet sources of such thick tuffs as the Rakosha, Gadareni, and Girish, the Novoselitsy tuffs, "lower tuffs" of Vyshkovo and Bergovskoye Kholmogor'ye, and the Cis-Carpathian tuffs, remain unknown. The same is true for lower Sarmatian tuffs, with the exception of Beregovskoye Kholmogor'ye, eastern Slovakia, and areas of Matra and Byukk, where rhyolite extrusives are present.

In establishing the position of individual volcanic layers, we naturally adhere to the present stratigraphic sections for the Trans- and Cis-Carpathias. However, as long as various correlations assign different positions to the same layers, we must consider the general patterns discernible in the several phases of magmatism, from the data covering a more extended area. More specifically, we believe it is more correct to associate the Danilovo tuffaceous layer with the lower Helvetian volcanism, although the onset of the explosions took place toward the close of the Burdigalian. By the same token, the "Novoselitsy dacite tuffs" should be moved from upper to lower Tortonian, as has been done lately by some students [7, 11], based on microfaunal data.

Only a few ideas on the geologic position of volcanics in the Trans-Carpathian Miocene are set forth in this paper. A more detailed treatment of this subject must be preceded by a petrographic and petrologic-chemical study of these rocks.

In conclusion, the author expresses his sincere gratitude to Professor V.S. Sobolev for his valuable advice; and to comrades I.V. Venglinskiy, V.N. Utrobin, L.A. Ivanova, and A.I. Klichuk, for their help in gathering the data.

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Received May 9, 1957

METASOMATIC FEATURES IN A DISTRICT OF THE KRIVOY ROG REGION [1]

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INTRODUCTORY NOTES

Metasomatic formations, described in numerous papers, are well developed in the northern Krivoy Rog basin [1, 7, 11]. As a rule, intensive replacement phenomena are observed in the favorable structural features associated with sharp bends and turns of the main Krivoy Rog syncline. We studied one such area for a number of years and were able to clarify certain features of the metasomatic process in that region.

Rocks of the middle and upper Krivoy Rog series are best developed in the area studied, where they are folded in an isoclinal syncline with flanks steeply dipping 75° to 80°. The main structure, especially in its centroclinal part, is complicated by numerous disharmonic steep folds of higher orders and by assorted breaks. Most important among the latter are dip-slip zones of faulting and shearing with associated feather fractures and steep zones of brecciation. This is a favorable condition for an upward movement of solutions and for an intensive thorough metasomatism of a large rock mass.

In most generalized form, replacement rocks present a horseshoe-shaped stock roughly corresponding to the main syncline. Its internal structure, however, is considerably complicated by the lithology and structure of the original rocks.

Identified among the metasomatic rocks of the area are magnetite-cummingtonite, magnetite-alkaline amphibole, hematite-hydrobiotites, aegirinites, talc-tremolite, tremolite-diopside, and other less common varieties. In addition, carbonate and siliceous varieties are well developed (Fig. 1).

Until recently, most authors, among them N. F. Anikeyeva [1], believed that metasomatism in the northern Krivoy Rog, and specifically in the subject area, had originated as a result of consecutive, commonly discontinuous processes, with the mineralogic features of individual metasomatic types resulting chiefly from different compositions of solutions participating in the several phases. A study of the metasomatic formations in the area studied has led to a different interpretation of the replacement process. The study was a structural-petrographic analysis of paragenetic associations of the replacement minerals. It is based on the following criteria: 1) the overall distribution and morphology of individual replacement rocks; 2) their relationship to the lithology and structure of the original regionally metamorphosed rocks; 3) relationship of the individual paragenesis of the replacement minerals among themselves; 4) chemical features of the individual paragenesis of replacement minerals; 5) the order of replacement of metasomatic minerals and their paragenesis by others; 6) the manner of alteration of minerals of variable composition.

As a result of the analysis of all these factors, we have come to the conclusion that all metasomatic formations in the subject area originated through a differential displacement of the substance of the original rocks after a prolonged action by a single metasomatic solution, without substantial addition of a juvenile component (with the possible exception of sodium). The diversity of metasomatic rocks, their composition, morphology, and spatial relationship are determined, not by the features of processes operative at various times, but rather by the lithology of the regionally metamorphosed original rocks, the mobility of the components, and the tectonic action on various segments. All of these factors affect the intensity of the metasomatism. Thus, in the areas of intensive folding, shattering, and shearing, as for instance in the centroclinal part, and in certain tectonic zones along the flanks, infiltrational replacement takes place in rocks subjected to a radical alteration of original composition and structure. In those places with a restricted circulation of

[1] Osobennosti metasomatoza v odnom iz rayonov severnogo grivorozh'ya.

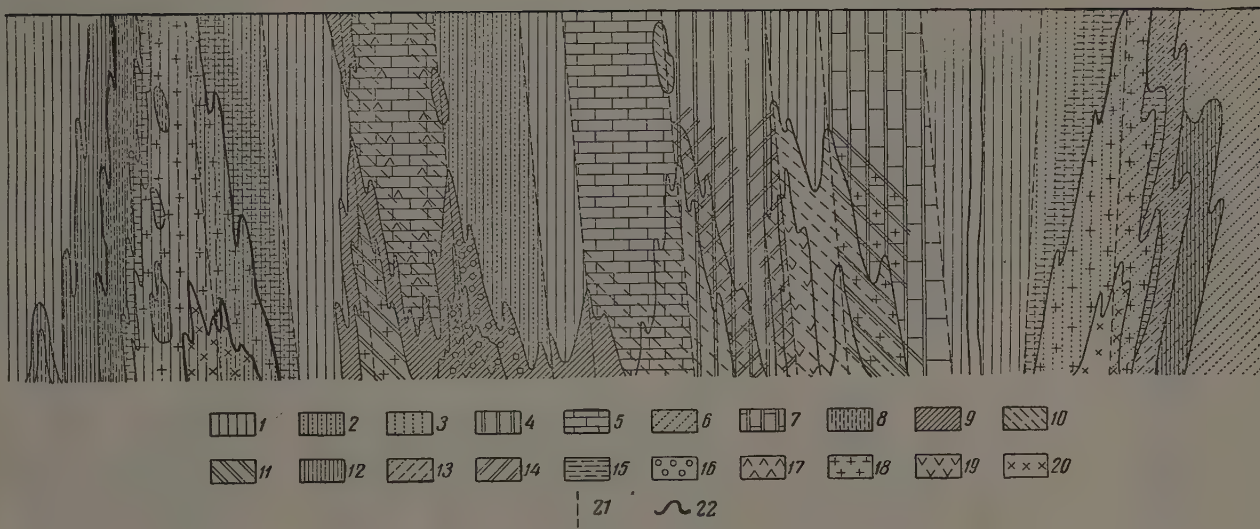

FIGURE 1. Outline of metasomatic development

1 -- Magnetite-biotite-quartz schists; 2 -- microcline-biotite-quartz schists; 3 -- mica schists; 4 -- quartzites; 5 -- dolomites; 6 -- ferrous quartzites and magnetite-chlorite-quartz schists; 7 -- quartz-mica schists; 8 -- crossite-rhodusite; 9 -- actinolite; 10 -- tremolite; 11 -- pargasite; 12 -- radiated amphiboles; 13 -- aegirine; 14 -- diopside; 15 -- hydrobiotite and chlorite; 16 -- epidote; 17 -- talc; 18 -- albite; 19 -- microcline; 20 -- quartz; 21 -- boundaries of original rocks; 22 -- boundaries of replacement rocks.

f solutions (such as the synclinal core), imetasomatic processes are observed wherein ecrystallization, without a substantial movelent of components, takes place at the conicts of chemically different rocks and in thologically homogeneous sequences. In both istances, local metasomatic zones originate, ith a definite and constant position for each ietasomatic-mineral paragenesis with relation) other parageneses and to the original rocks. uch a structure may be explained by D. S. orzhinskiy's hypothesis of differential mobily of the components, in the process of eplacement, and of the formation of a metaomatic zonation [6].

Inasmuch as the composition of replacement iinerals, in our opinion, is affected chiefly y the lithology of the regionally metamorhosed original rocks, a detailed stratigraphy f the area will not be given but only its main ithologic groups will be listed. They are as ollows: 1) magnetite-biotite-quartz and quartz-iotite schists, with less common garnetuartz-biotite schists ($MgO-FeO-Al_2O_3-SiO_2$); ?) ferrous quartzites and magnetite-chloriteuartz schists ($FeO-SiO_2$ and $MgO-FeO-SiO_2$); 3) sericite-quartz, less commonly garnetsericite-quartz, schists in places with chlorite or biotite ($Al_2O_3-SiO_2$); 4) mica microclineuartz schists ($Al_2O_3-SiO_2$) with a high K_2O content; 5) quartzites; 6) dolomites (carbonate rocks with a high dolomite content).

Magnetite-biotite-quartz and quartz-biotite schists are developed chiefly along the western limb, and in the centroclinal and core segments. They are marked by an alternation of beds relatively rich (ferrous layers) and relatively poor (schist beds) in iron, which is reflected in the accepted stratigraphic sequence (2). Ferrous quartzites, also a component of the ferrous layers, are developed chiefly along the western limb. Sericite-quartz schists with chlorite, biotite, and some intercalations of mica-microcline-quartz schists lie in a consistent band on both limbs and in the centrocline. It is to be noted that they have undergone a thorough metasomatic transformation and are present as isolated relicts in the northern part of the area where the metasomatism was not as intensive. The core of the syncline is formed by a complex alternation of dolomites and quartzites, with assorted schists whose amount increases in the upper part of the section.

All of these rocks, which belong to the middle and upper divisions of the Krivoy Rog series, rest conformably, commonly with gradual transitions from one to another, determining the presence of intermediate rocks such as carbonate-schistose rocks, carbonate quartzites, etc.

INFILTRATIONAL METASOMATISM

Infiltrational metasomatism is most intensively developed in association with zones of dip-slip faulting and shearing along the synclinal limbs and with numerous tectonic complications in the centrocline. These breaks are present in magnetite-biotite-quartz and sericite-quartz schists and, to a smaller extent, in ferrous quartzites. Each of these rocks gave rise to sections with a definite number of replacement zones with mineralogic compositions of their own.

In each specific case, the boundaries of individual metasomatic zones do not present ideal concentric surfaces. They form, instead, a very complex network. A study of the morphology of individual associations of metasomatic minerals is necessary for the establishment of their genetic relationship. We propose the following main morphologic types of metasomatic minerals:

1. The zonal form. This is a stratified sequence of a definite paragenesis of metasomatic minerals, between the original rock and the next metasomatic zone; or between two different zones, wherein it replaces the first zone and is replaced by the second.

2. Frontal, bedding plane, or cutting veins of metasomatic minerals in the original rocks or in advanced metasomatic zones. These may be regarded as more or less elongated extensions of a given metasomatic zone, related to weakened segments (cleavage and shearing fractures, etc.) which facilitated the advance of solutions ahead of the main replacement front (the zonal form). When such a vein cuts several zones, zonation appears along its contact, which repeats the main zonation, on a small scale.

3. Veinlike relicts of the preceding zones occur in the following ones. In extreme instances of replacement, the suturetype relict seams of minerals from the preceding zones are formed. A cursory study of such replacements may lead to the conclusion of a later differentiation of vein minerals, with reference to the enclosing rock.

The subject area is characterized by a banded replacement pattern, with an alternation of bedding frontal and relict veins, resulting from the rock schistosity. In vertical cross section these formations have a comblike pattern.

4. Veins of mineral associations from the preceding zones are commonly developed as a breccia cement. As a rule, they occur in brecciated zones of banded replacement rocks and are made up of minerals of the enclosing rocks. For instance, in rocks presenting an

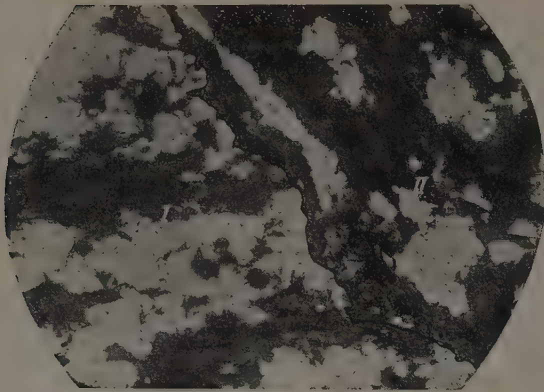

FIGURE 2. Breccialike albitite.

I -- Fragment of banded crocidolite albitite; II -- albite-crocidolite breccia cement. 10X, without analyzer.

alternation of consequent "zonal" alkaline-amphibole, hematite-hydrobiotite, and albite associations, the breccia cement consists of the same minerals, i.e. alkaline amphibole, hematite, hydrobiotite, and albite (Fig. 2). There are instances of albite being cut by alkaline-amphibole and hematite-hydrobiotite veins which were commonly assigned, for that very reason, an independent and later origin. On the basis of the great mineralogic similarity of these veins and the enclosing rocks, we offer D.S. Korzhinskiy's [5, 6] concept of supersaturation of metasomatic solutions in pores as an explanation for such formations. Upon entering the hollows which originate in the metasomatic process, these solutions precipitate minerals made up of the components of the enclosing rocks. D.S. Korzhinskiy believes that colloidal reactions play a considerable part in this process. Indeed, collomorphous, fibrous, and spherulitic structures are fairly common in the cement of breccialike albitites, along with the development of crocidolite which, in the opinion of many authors, is of colloidal origin [13, 14]. The same origin may be assigned to veins of a mineral or an association of minerals in rocks of the same composition, such as crocidolite veins in alkaline-amphibole rocks, and comblike albitites in schistose albitites.

Metasomatic columns of type one are developed in quartz-biotite, magnetite-biotite-quartz and to a smaller extent in garnet-biotite-quartz schists. Accessory minerals are represented by tourmaline, apatite, zircon, and sphene. Chemical analyses of quartz-biotite schists (analyses 1, 2) show a considerable preponderance of alumina over alkalis, of ferrous over ferric iron, and of K_2O over Na_2O. Ratio $(Na_2O + K_2O)/Al_2O_3$ is equal to 0.45, in analysis one; to 0.23, in analysis two; ratio K_2O/Na_2O is 4.3 and 16.0, respectively. (Here, as elsewhere in this paper, ratios of molar % are given).

The first metasomatic rocks are magnetite-cummingtonite schists which we have named the "amphibole zone." First, radiated amphiboles are developed selectively in layers enriched with biotite and magnetite. As the metasomatism progresses, garnet and quartz are replaced and the rock becomes coarser. Light-green hydrobiotite and magnetite occur in the same paragenesis with radiated amphiboles. These minerals cannot be regarded as relicts of magnetite and biotite of the original rocks because of the different manner of their occurrence and distribution. Furthermore, where cummingtonite shows obvious replacement structures, as compared with biotite and magnetite of the original rock, its relation with hydrobiotite and magnetite of the amphibole schists is the same as with the contemporaneous metasomatic minerals. It is true that, because of the low mobility of most of the components, the amphibole zone exhibits a direct relation between the magnetite and biotite content in the original regionally metamorphic schists, on one hand, and the magnetite and hydrobiotite content in cummingtonite schists, on the other.

Cummingtonite, which accounts for 70 to 80 percent of amphibole schists (γ = 1.670-1.682; α = 1.648-1.655; 2V ≈ + 80°;

Table 1

Chemical Analyses of Schists in the Northern Krivoy Rog

Oxides	Analyses (% by weight)													
	1	2	3	4	5	6	7	8	9	10	11	12	13	14
SiO_2	53,58	55,13	63.78	30,03	45,98	51,58	45,32	51,16	49,80	54,40	54,33	47,43	45,07	53,48
TiO_2	0,62	0,43	Not det.	0,04	0,24	0,21	0,17	0,56	—	—	0,01	0,78	0,79	—
Al_2O_3	13,45	17,37	5,26	1,11	5,44	4,08	7,71	2,88	1,56	2,90	6,71	4,71	3,47	5,32
Fe_2O_3	5,49	1,24	0,86	31,22	4,92	0,18	2,23	14,41	18,62	14,40	17,81	20,18	25,61	15,16
FeO	13,12	12,41	25,83	27,46	32,86	29,97	25,40	10,13	10,59	3,80	—	10,00	10,03	3,44
MnO	0,02	0,20	—	—	—	—	—	—	—	—	—	—	—	—
MgO	2,43	3,24	2,85	5,04	7,39	7,45	5,40	12,48	9,30	15,50	11,82	7,64	5,78	10,90
CaO	0,60	0,31	0,23	1,83	0,63	2,73	0,21	5,20	0,45	1,50	0,69	1,84	1,17	0,72
Na_2O	0,65	0,10	0,39	0,86	—	0,72	1,11	2,14	8,80	6,50	4,82	4,48	5,61	6,30
K_2O	4,65	3,48	0,44	0,37	0,19	0,12	2,60	0,12	traces	—	0,80	0,22	0,17	0,70
Loss on ignition	5,00	5,85	0,11	1,52	2,55	2,61	7,23	1,29	0,65	0,80	2,30	2,15	1,48	2,32
H_2O	0,28	0,40	0,08	0,32	0,22	0,21	2,57	0,09	—	—	0,47	—	0,10	0,94
Total	99,89	100,16	99,83	99,80	100,42	99,86	99,95	100,46	99,77	99,80	99,76	99,43	99,28	99,28

Note: comma represents decimal point.

1 -- After Yu. Ir. Polovinkina; 2, (1), 3 -- cummingtonite-biotite-quartz schist; M.I. Makarova, analyst; 4 -- composite specimen of magnetite-cummingtonite schists, after V.I. Sinyakova; 5 -- cummingtonite, after Yu. R. Polovinkina; 6 -- cummingtonite; M.I. Makarova, analyst; 7 -- light-green hydrobiotite from amphibole schists, after A.P. Nikol'skiy; 8 -- M.I. Makarova, analyst; 9 (15), 10 (16); 11 -- after A.P. Mikol'skiy; 12, 13, 14 -- M.I. Makarova, analyst;

dispersion 2V−r < v), increases in grunerite molecule content on approaching the rear part of the zone (γ = 1.702; α = 1.668; 2V = 90−80°; dispersion 2V−r > v). Hydrobiotite has the following optical properties: γ = 1.618−1.630; α = 1.591−1.603; 2V = +5−7°.

A comparison of the analyses of cummingtonite schists and their component minerals with the analyses of original rocks (see Table) reveals, first of all, a marked change in the alkali ratio. Where there was a definite preponderance of K_2O over Na_2O in quartz-biotite schists, their ratio is reversed in magnetite-cummingtonite schists where it is 0.3 on the average (analysis 4). It is significant that the change in the K_2O/Na_2O ratio, with Na_2O increasing, takes place immediately after the appearance of cummingtonite (analysis 3). This change in the alkali ratio, along with the leaching of K_2O, evidently points to a change of K_2O and Na_2O to a mobile state, more so for K_2O.

The second feature of cummingtonite schists is the accumulation of MgO and FeO,

Table 1, continued

Chemical Analyses of Schists in the Northern Krivoy Rog

Oxides	Analyses (% by weight)						
	15	16	17	18	19	20	21
SiO_2	53,54	44,60	47,98	51,56	48,67	46,00	38,22
TiO_2		—	0,53	0,48	0,06	traces	0,66
Al_2O	0,46	11,21	15,42	13,92	3,91	7,34	8,07
Fe_2O_3	20,06	9,78	6,21	8,96	3,87	3,00	33,18
FeO	4,10	12,04	10,59	8,73	24,15	20,64	7,09
MnO	—	—	—	—	—	—	—
MgO	12,47	7,73	6,25	5,48	3,47	7,79	4,89
CaO	0,27	4,33	2,40	0,23	3,56	2,88	2,93
Na_2O	5,95	3.17	4,54	3,19	1,20	1,58	2,68
K_2O	0,46	0,65	0,79	1,48	2,09	1,54	0,16
Loss on ignition	2,65	5,27	3,98	4,07	6,28	6,83	1,98
H_2O	0,19	0,92	0,88	1,50	2,54	2,58	0,48
Total	100,15	99,70	99,57	99,73	99,80	100,18	100,34

Oxides	Analyses (% by weight)						
	22	23	24	25	26	27	28
SiO_2	52,94	39,02	56,41	65,00	55,56	59,30	51,64
TiO_2	0,92	0,32	0,36	0,44	0,53	0,25	—
Al_2O	11,27	4,12	13,07	16,13	8,48	8,56	1,13
Fe_2O_3	7,19	33,65	11,34	3,16	9,37	5,21	0,05
FeO	9,90	10,70	5,67	2,39	9,68	8,17	33,88
MnO	—	—	—	—	—	—	—
MgO	6,27	3,84	1,62	1,62	6,83	9,37	8,40
CaO	4,05	1,07	0,80	0,52	0,88	1,24	0,97
Na_2O	3,72	3,34	7,38	10,12	6,03	6,11	0,83
K_2O	0,22		1,05	0,00	0,00	0,55	0,06
Loss on ignition	2,84	1,21	1,49	0,73	1,02	1,42	0,10
H_2O	0,68	3,04	0,61	0,16	0,14	1,32	—
Total	100,00	100,00	99,80	100,27	99,42	100,50	100,64

Note: comma represents decimal point.

15 (11), 16, 17 -- M.I. Makarova, analyst; (19), 18 -- hydrobiotite from albite, north Krivoy Rog, after Yu. Ir. Polovinkina; 19 -- hydrobiotite from albitite, northern Krivoy Rog, after Yu. Ir. Polovinkina; 20 -- hydrobiotite from an albite vein, Third Internationale Mine, middle Urals, after V.A. Zavaritskiy (3); 21 -- M.I. Makarova, analyst; 22 -- analysis 21 converted for hydrobiotite; 23 -- average analysis of 9 albitized schists specimens, after V.I. Sinyakov; 24 -- average analysis of 7 albite specimens, after V.I. Sinyakov; 25 -- glaucophane albitite, L.P. Orlova, analyst; 25 -- M.I. Makarova, analyst; 27, (17), 28 -- M.I. Makarova, analyst.

chiefly in radiated amphibole. Concentration of magnetite is commonly found in the rear part of the zone, where Fe_2O_3 also increases (analysis 4). Typical for cummingtonites, developed on quartz-biotite schists, is a high alumina content, as much as 5.44 percent (analysis 5).

Magnetite-cummingtonite schists, whose thickness reaches 20 to 30 m, have a much wider development than the other metasomatic rocks of type one. This is so because they, being an advanced metasomatic zone, require but an insignificant change in the chemical composition of original rocks. Even a small infiltration of solutions results in replacement of quartz-biotite schists by the magnetite-cummingtonite. However, they are typically metasomatic products. This is suggested by their association with weakened and sheared

ones; their lenticular form, locally trans-ersal to the enclosing rocks, and the pres-nce of cummingtonite veins; the replacement f minerals of the original rocks by cumming-nite and by the presence of coarse-grained extures; their close association with the llowing (spatially) metasomatic rocks; the hemical features of magnetite-cummingtonite ocks, and above all the behavior of the lkalis.

The next metasomatic zone, from 1 to 12 n thick, which replaces the amphibole zone, s made up of the products of the paragenesis f magnetite and alkaline amphiboles of rossite-rhodusite type. A regular change in he composition of alkaline amphiboles and in he relative amounts of rock-making minerals akes place across the strike. Magnesium-errous varieties are developed in the fore art of the zone where they pseudomorphously eplace the radiated amphiboles. In the rear art, they give place to fibrous ferrocrocido-ites. A similar variability of alkaline amphi-oles was noticed by a number of authors for imilar manifestations of alkaline metasomat-sm in ferrous quartzites [8, 10].

Alkalinization of cummingtonite begins with he appearance of blue fringes with a pleo-hroism in light blue-lilac-yellow hues. For he same optical orientation of cummingtonite, he refraction in these fringes decreases to ., 650 – 1, 660. The low-alkali content and the redominance of ferrous over ferric iron in uch amphiboles (Fe_2O_3:FeO = 0.64) definitely oints to a considerable cummingtonite con-ent. However, the overall composition and he ratio of total alkalis to alumina, 1.56 analysis 8), correspond to a true alkaline mphibole of the crossite-rhodusite type [12].

A conversion of the analysis gives the ollowing crystallochemical formula:

$$(Na_{0.58}K_{0.18}Ca_{0.95})_{1.71}(Mg_{2.62}Fe_{1.20}Fe_{1.54})_{5.39}[OH]_{1.24}$$

$$[Al_{0.48}Ti_{0.06}Si_{7.56}O_{22.76}].$$

The first fully individualized variety of alkaline amphibole has an unusual orientation: b = γ, with a approaching α. 2V = -40-50°, and with sharp dispersion v > r. In sections |(010), β = 38° for λ = 578mμ and 17° for λ = 436mμ, i.e. the horizontal dispersion[1] is 21°; γ = 1.664; α = 1.652 (λ = 578). An amphibole from northern Krivoy Rog, with similar optical properties is described by Yu. Ir. Polovinkina [11].

The further alteration of amphiboles is expressed in higher color and in stronger refraction. The distribution of pleochroic colors along the crystallographic directions is preserved (c - greenish blue; b - lilac-violet; a - yellow) and only their absolute intensity changes. The refractory indices grow at different rates in different directions, changing the orientation of the ellipsoid. Refraction increases faster along axis a and the orientation becomes normal: β coincides with b and γ approaches a. A simultaneous decrease in the extinction angle and the bisetrix dispersion takes place. As an intermediate variety, an amphibole with the following properties may be cited: b = γ, a $\approx \beta$, c $\approx \alpha$; c : α for λ 578 = 12°, cα for λ 436 = 4° (horizontal dispersion, 8°); 2V $\approx$ 50° with dispersion v > 4; γ = 1.672, α = 1.660 (λ = 578mμ).

Amphiboles with such properties have been described in many instances. The closest to our variety is bababudanite from Mysore (analysis 9), rhodusite in the glaucophane-riebeckite series from the Kaoko Veld, South West Africa (analysis 10).

With the uneven growth of refraction indices along axes a and b, it is only natural that a variety should originate, with 2V = 0. We have not come across any strictly mono-axial varieties. Those amphiboles in which we found a minimum axial angle of 15° to 20° had the following optical properties: b = β, a $\approx \gamma$, c $\approx \alpha$; c : α for λ 578 mμ = 23° and for λ 436 mμ = 21°; γ = 1.676, α = 1.6665. The possibility of a monoaxial variety among alkaline amphiboles was voiced by N. I. Nakovnik [3].

Typical varieties of pseudomorphous alkaline amphiboles with a normal orientation have γ = 1.673–1.680; α = 1.668–1.674; γ–α = 0.008–0.010; c : α = 12–24°; 2V $\approx$ -60° with dispersion of the optical axes angle v > r = 4° and the bisectrix dispersion v < r = 6°. Judging by the optical properties, the nearest to the described variety is riebeckite from amphibole-magnetite ores of northern Krivoy Rog, with γ = 1.680 and α = 1.670 (analysis 11). On the basis of our analysis, it also should be assigned to crossite-rhodusite.

Crocidolites of the rear part of the zone also display various optical properties depending on the intensity of coloring, which in turn is related to the iron content. Refraction in crocidolites varies in the range, γ = 1.685–1.700; α = 1.670–1.696; γ–α = 0.007–0.004; c : α = 10–12°, and is nearly zero for the most intensely colored varieties, with a low birefringence. Oblique dispersion is 5° to 3°. The conoscope reveals negative biaxial crystals with a sharp dispersion v > r. Analysis 12, presents an "intermediate" crocidolite

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[1] Elsewhere in this paper, dispersion is given for λ 578–436 mμ.

with $\gamma = 1.687$, $\alpha = 1.681$; $c : \alpha = 8^{\circ}$, analysis 13, the most intensely colored fine fibered with $\gamma = 1.698$, $\alpha = 1.692$ and $c : \alpha = 1-2^{\circ}$. A conversion of the analyses give the following crystallographic formulae:

$$(Na_{1.3}K_{0.04}Ca_{0.30})_{1.64}(Mg_{1.7}Fe_{1.25}Fe_{2.26})[OH]_2$$

$$[Al_{0.82}Ti_{0.09}Si_{7.09}O_{21\ 84}(OH)_{0.16}]$$

$$(Na_{1.68}K_{0.04}Ca_{0.19})_{1.91}(Mg_{1.34}Fe_{1.30}Fe_{2.26})_{4.90}$$

$$[Al_{0.62}Ti_{0.09}Si_{7.01}O_{22.46}].$$

Crocidolites known from published data usually are either more ferrous [14] or richer in MgO [4, 13] than those investigated here. The closest to our crocidolites is blue crocidolite-asbestos from Robertstown, South Australia [13]. From the chemical analyses of various recognized alkaline amphiboles, we deem it possible to assign our crocidolites to crossite-riebeckites. As compared with crossite-rhodusites, crossite-riebeckites contain less SiO_2 and MgO and more of Fe_2O_3 and Na_2O. It is of interest that the ferrous iron content is nearly the same for all alkaline amphibolites of the area: 10 to 12 percent. Ratio $(Na_2O + K_2O)/Al_2O_3$ rises to 2.7 as against 1.56 for crossite-rhodusite, with Fe_2O_3/FeO increasing from 0.64 to 0.94, both changes corresponding to the sodium increase. This corroborates Yu. Ir. Polovinkina's conclusion that the degree of oxidation of iron depends on the alkalinity of the medium. However, this is true only for silicates, because magnetite in alkaline-amphibole rocks shows no evidence of martitization. Alkalinity here means the alkali-alumina ratio rather than their absolute content [12]. Depending on the concentration of alkali, alumina enters either the cation or anion fraction of the silicates. With a high-alkali content in amphiboles, $Al^{\cdot\cdot\cdot}$ isomorphously replaces $Si^{\cdot\cdot\cdot\cdot}$ in the anion fraction, with $Fe^{\cdot\cdot\cdot}$ taking the place of a trivalent cation in sixfold coordination. Furthermore, insofar as the Mg and Ca content in amphiboles decreases with an increase in Na, $Fe^{\cdot\cdot}$ changes to $Fe^{\cdot\cdot\cdot}$ for valent compensation. It is significant that ferric iron considerably exceeds ferrous iron in alkaline amphiboles with a low-alumina content (analysis 15).

The conclusion of Yu. Ir. Polovinkina and A.P. Nikol'skiy to the effect that the iron content decreases considerably where alkaline amphiboles replace the radiated amphiboles, with the accompanying decrease in magnetite, is also confirmed by our data (in % of weight):

Iron oxides	Cummingtonites		Crossite-rhodusites	Crossite-riebeckites	
Fe_2O_3	0.18	4.92	14.41	20.18	25.61
FeO	29.97	32.53	10.13	10.00	10.03
Total iron, as converted into FeO	30.13	37.02	23.10	28.35	30.04

Petrographic observations fully agree with chemical analyses. Indeed, the replacement of cummingtonites by crossite-rhodusites is accompanied by a removal of magnetite. Toward the rear of the zone, on the other hand, the increasing iron of the amphiboles leads to the amphiboles replacing the magnetite until the latter disappears completely in crocidolites with an overall iron content of about 30 percent.

Analyzed for iron, specimens of crossite-rhodusite-magnetite ores from the front part of the zone were found to have 44.31 percent of Fe_2O_3 and 29.30 percent FeO, i.e., a ratio of 0.68. In crocidolite analyses 12 and 13, this ratio is 0.91 and 1.04 respectively. It follows that, on the whole, the degree of oxidation of iron increases across the zone strike, but the Fe_2O_3/FeO ratio does not exceed one. Evidently, this is responsible for the lack of martitization of magnetite in its replacement by crossite-riebeckite.

The hematite-alkaline-amphibole zone is followed by the hematite-hydrobiotite. This zone, located between crocidolite schists and albitites, is usually thin – from microscopic to a few tens of centimeters, with a widely developed bedding plane and cutting hematite-hydrobiotite veins in magnetite-alkaline-amphibole rocks. Because of this, the development zone of this association reaches several meters.

The peculiar morphology of the hematite-hydrobiotite association is apparently the result of the continual fracturing of the magnetite-alkaline-amphibole zone, in the metasomatic process. It may have been a result of incessant shifting under the condition of a lower competency of alkaline-amphibole schists as compared with amphibole schists and albitites. In addition, fracturing may have been caused by the volume change in metasomatic replacement of high-iron rocks by albitites.

It should be noted that the hematite-hydrobiotite zone may be missing locally, with crocidolite replaced directly by albite. Its absence may be attributed to the composition of the original amphibole: high-alumina amphiboles of a glaucophane type probably

are directly replacable by albite. Most likely, however, the development of this zone is associated with the specific condition of fracturing and formation of hollows at the boundary between alkaline-amphibole rocks and albitites. This is suggested by the veinlike forms of hematite-hydrobiotite bodies and by the absence, commonly, of a hydrobiotite zone about the thin veinlets of crocidolite in albitites where the formation of hollows is difficult to conceive. Another indication of the precipitation in cavities is the coarse crystalline, spherulitic, locally collomorphous bodies of hematite-hydrobiotites and the presence of minerals with a high content of volatile components (CO_2 in carbonates, H_2O in hydrobiotite and malacon, F in apatite, B in tourmaline).

However, this hematite-hydrobiotite association is undoubtedly a zone of our metasomatic column rather than an independent generation. The association is everywhere located between the magnetite-alkaline-amphibole and albite zone and it replaces crocidolite schists, and is in turn replaced by albitites.

The mineralogic composition of this zone is represented by hydrobiotite, hematite, calcite, sphene, malacon, apatite, less commonly by tourmaline. Hydrobiotite forms a number of varieties, all with direct extinction and positive elongation, with $2V = 0\text{-}4^{\circ}$. Coloring along γ changes from brown-green, at times with a reddish tint, to light grass-green; along α, from orange to light yellow, nearly colorless. For intensely colored varieties, $\gamma = 1.600\text{-}1.583$, $\alpha = 1.566$ (1.554, $\gamma\text{-}\alpha = 0.035$; for pale varieties, $\gamma\text{-}1.579\text{-}1.570$, $\alpha = 1.599\text{-}1.549$, $\gamma - \alpha = 0.020$. In brown-red and brown-green varieties, γ is as high as 1.620-1.630. Data close to ours are cited for Krivoy Rog biotites [11] and for hydrobiotites from pyrite deposits in the Middle Urals [3].

Analysis 16 represents an intensely colored variety with $\gamma = 1.162$ and $\alpha = 1.578$, and analysis 17 is of a lighter-colored hydrobiotite with $\gamma = 1.582$ and $\alpha = 1.554$. For comparison, three analyses from published data are given for hydrobiotites, which on the whole are fairly similar to our analyses (analyses 18, 19, 20).

Conversion of analyses gave the following crystallographic formulae:

Analysis 16

$$(Na_{0.48}K_{0.08}Ca_{0.34}[H_3O]_{0.12})\,(Mg_{0.84}Fe^{\cdot\cdot}_{0.73})_{1.57}$$
$$(Al_{0.22}Fe^{\cdot\cdot\cdot}_{0.54})_{0.76}[OH]_2 \quad X$$
$$[Al_{0.75}Si_{3.25}O_{10}] \cdot 0.34 \cdot H_2O;$$

Analysis 17

$$(Na_{0.60}K_{0.10}Ca_{0.17}[H_3O]_{0.13})\,(Mg_{0.65}Fe^{\cdot\cdot}_{0.62})_{1.27}$$
$$(Al_{0.60}Fe^{\cdot\cdot\cdot}_{0.33})_{0.93}[OH]_{2.07} \; X$$
$$[Al_{0.66}Si_{3.34}O_{10}] \cdot 0.1H_2O$$

A comparison of analyses of crocidolites and hydrobiotites reveals that the latter have a considerably higher Al_2O_3 content and lower amounts of Na_2O and Fe_2O_3, with FeO, CaO, and MgO remaining approximately at the level of ferrocrocidolites. It follows that the main chemical feature of the replacement of crocidolite by hydrobiotite is the displacement of iron oxide by aluminum from the cation fraction of silicates, the Fe_2O_3/FeO ratio decreasing sharply to 0.36-0.26 as compared with 0.91-1.04 in crocidolites. Excess iron is precipitated as hematite.

The increase in alumina and the decrease in Na_2O, which is partially displaced by oxonium, also leads to a sharp decrease in the Na_2O/Al_2O_3 ratio which is 0.56-0.65 in hydrobiotites as against 1.6-2.7 in crocidolites. Where, in formulae for crocidolites, atomic factors for Na exceeded those for Al, fully entering the anion radical, the situation is reversed in hydrobiotites. In addition, aluminum here, too, enters the anion fraction. Thus, as far as the ratio of alkalis to alumina is concerned, an acid environment prevails in the hematite-hydrobiotite zone.

Analysis 21 represents a hematite-hydrobiotite rock with an average content of 20 percent hematite and 80 percent hydrobiotite. For control, it is converted for hydrobiotite content (analysis 22), which is very similar to the analyses of pure hydrobiotite fractions. This hematite-hydrobiotite rock (analysis 21) directly replaces crocidolite schists (analysis 13). A comparison of the analyses reveals a sharp increase in Al_2O_3 (by the factor of two) and FeO in the hematite-hydrobiotite rock, and a decrease in Na_2O (2.1 times). The total amount of iron remains nearly the same. Converted into ferrous iron, its amount in crocidolite schists is 33.04 percent, with 36.89 percent in hematite-hydrobiotite rocks. Ratio Fe_2O_3/FeO is 1.0 and 2.1, respectively, with the bulk of ferric iron, in the latter, tied up in hematite, inasmuch as the ratio for hydrobiotite is but 0.33. Consequently, in the replacing of crocidolites by a hematite-hydrobiotite aggregate, iron in the silicate lattice is displaced by aluminum and is precipitated as hematite, with its overall amount remaining approximately the same.

Malacon (Fig. 3), sphene (Fig. 4), and, less commonly, tourmaline concentrate in the hematite-hydrobiotite zone where they

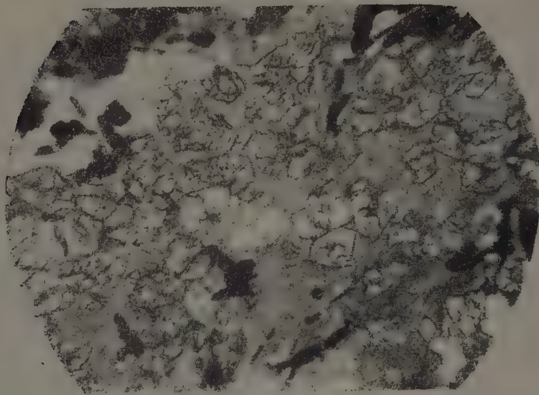

FIGURE 3. Malacon in a hematite-hydrobiotite aggregate. 40X; without analyzer.

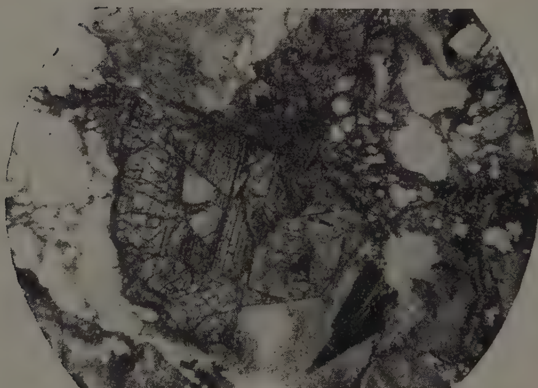

FIGURE 4. Sphene in an albitized hydrobiotite aggregate. 10X; without analyzer.

commonly form large accumulations. Apatite is also fairly common. Their precipitation is probably connected with specific features of the formation of this paragenesis in fractured zones of lower pressure.

The next metasomatic zone is made up of albitites, from several to a few tens of meters thick. The fore part of the albitites contains numerous layers and relicts of the preceding parageneses, chiefly of the hematite-hydrobiotite and magnetite-alkaline-amphibole zones.

Albite is the principal mineral of this zone, whose paragenesis includes glaucophane, hydrobiotite, magnetite, pyrite, all in definitely subordinate amounts. Albitites are usually characterized by a two-mineral paragenesis, either albite-glaucophane or albite-hydrobiotite. The content of dark-colored minerals decreases toward the central part of the zone.

Several structural varieties occur among albitites, commonly regarded as generations. We recognize the following varieties: hypidiomorphic, blastic, comblike, checkerboard, and vein albitites. Breccialike albitites present a structural variety of their own. Hypidiomorphic and rosette-shaped albitites are developed in the frontal part of the replacement zone; blastic albitites make up the central and rear parts where their growth

as stunted. Comb and checkerboard albitites re definitely associated with shearing zones. he comb albitites originate in the cyrstalliza-on of mylonitized albititites. The checker-oard albitites, located as a rule in a sheath bout the comb, are associated with the ecrystallization of strongly fractured zones. ıdeed, in fractured zones of cataclastic rock f various grain size, and in the presence of bundant pore-filling solutions, the latter may e undersaturated with regard to finer grains, ith a high surface energy, and supersaturated ith regarded to coarser rocks. In such cases, oarser grains -- crystals of checkerboard nd comb albite here -- will grow at the xpense of the finer ones [6].

The above-named varieites of albitites are ynchronous. Some vein albitites with distinct ringes may belong to a later hydrothermal rocess.

The chemical composition of all albite arieties is very consistent, as shown by hemical analyses and microscopic study, and s that of albite with 3 to 4 percent Na. This bviously points to a rather high sodium otential, as compared with potassium, inas-nuch as the latter is nearly fully displaced rom the Al-silicate lattices.

Hydrobiotite and glaucophane of albitites ıre transitionally related with hydrobiotites ınd alkaline amphiboles of the preceding ones, at the same time retaining their own omposition and properties. Glaucophane in ılbitites is developed in radiated, filiform, ınd acicular bodies with a very high Al_2O_3 ınd decreased Fe_2O_3 content (analysis 26). eO remains at the level with amphiboles of he area, namely 9.68 percent. The crystal-ographic formula for glaucophane is:

$$Na_{1.68}Ca_{0.14})_{1.829}Mg_{1.47}Fe^{\cdot\cdot}_{1.17}Fe^{\cdot\cdot\cdot}_{1.03}Al_{1.42})_{5.09}$$

$$[OH][Si_{8.05}Ti_{0.08}O_{23}]\,.$$

t is significant that all of the aluminum in glaucophane is tied up in the cation fraction. Ratio Na_2O/Al_2O_3 is close to one (1.18); accordingly, glaucophane is nearly "neutral." However, ratio Fe_2O_3/FeO drops sharply to).44 as compared to 1.04 in crocidolites, and he overall iron content decreases to 18.11 percent (converted for ferrous). In its chemi-cal composition and optical properties, our glaucophane comes closest to that from the ısland of Hokkaido (analysis 27).

Besides the silicates, albitites carry mag-netite and pyrite. Magnetite originates in small idiomorphic crystals along the replace-ment front of advanced zones and causes muschketowitization of hematite in the hematite-hydrobiotite zone.

Along with ferromagnesian minerals, sphene (Fig. 4) and malacon undergo replace-ment. Apatite is preserved in albities, but in considerably smaller amounts as compared with the hematite-hydrobiotite zone. Generally speaking, apatite is a "transient" mineral of the metasomatic column, occurring in all zones and in amounts comparable with its content in original rocks. This points to a very low mobility of phosphorus in the meta-somatic process. However, some concentra-tion of apatite in the hydrobiotite zone is observed.

A comparison of chemical analyses for albitites with those for the preceding zones reveals a large increase in SiO_2, Al_2O_3, and Na_2O in the former, and a decrease in FeO, MgO, and CaO (analyses 23, 24). Typical for albitites is the ratio Na_2O/Al_2O_3, approaching one; and Fe_2O_3/FeO, less than one (0.9-0.6). This establishes the environment as "neutral" and reducing, which is corroborated in the formation of magnetite, muschketowite, and pyrite in albitization. Petrographic study shows that Mg, Fe, Ti, and Zr minerals are unstable under such conditions.

The last metasomatic zone is the quartz zone. Macroscopically, metasomatic quartzites strongly resemble albitites, which renders them difficult to map. Quartzites possess all the features of a metasomatic rock, such as the lack of clean-cut fringes, replacement of albitite minerals by quartz, and albitite relicts in quartz. Textures in all preceding zones were determined by the reverse iso-morphism rule, i.e., that succeeding minerals (spatially) are more idiomorphic than the pre-ceding. Quartz, which is represented in quartzites by coarse xenomorphic zones, "idiomorphically dissolves" the albite. Thus, poikiloblastic quartz with numerous idiomor-phic little crystals of albite originates in the replacement of blastic albitite (Fig. 5).

The chemical environment of the quartz zone is determined by the mobility and leach-ing of all components and by accumulation of excess SiO_2.

To the second type of infiltrational meta-somatism we assign columns developed in ferrous quartzites and in chlorite-magnetite-quartz schists, i.e., in rocks with a marked preponderance of FeO-SiO_2 and MgO-FeO-SiO_2. A typical feature of these columns is the presence of an aegirinite zone correspond-ing in its position to the magnetite-alkaline-amphibole zone of the type one columns.

In its pure form, the ferrous quartzite column is simple, consisting of 1) magnetite-quartz; 2) aegirine; 3) quartz.

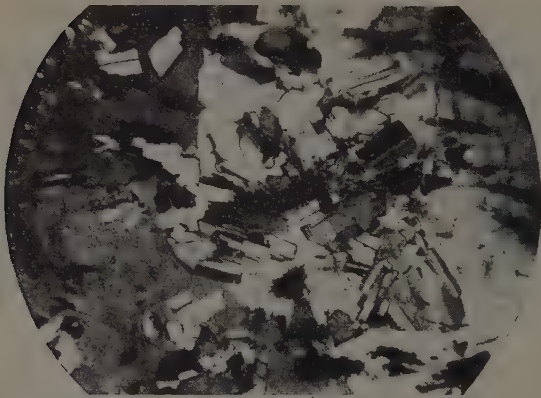

FIGURE 5. Idiomorphic relicts of albite in quartz replacing the albite. 40X; crossed Nicols.

In the metasomatism of chlorite-quartz schists, with chlorite present as a ferro-magnesian variety of aphrosiderite, a cummingtonite zone originates first. In contrast to the cummingtonite of type-one columns, the cummingtonite which originates at the expense of chlorite-quartz schists, is very low in alumina (analysis 28).

Alkalinization of cummingtonite is observed in front of the aegirinization zone, with an accompanying slight precipitation of magnetite. This process, however, is of a quite different magnitude from the precipitation of magnetite in type-one columns, because of the nearly equal iron content in both cummingtonite and aegirine.

The mechanism of formation for type-two column is fairly simple. The infiltration of sodium solutions ties up the iron in aegirine and oxidizes it to a trivalent form. In this process, magnesium becomes relatively more mobile, which leads to the formation of a cummingtonite zone. In further infiltration of solutions, iron changes to a mobile form, with the accompanying quartzitation of aegirinites.

A direct quartzitization of aegirinites, however, seldom occurs. In most instances, they are replaced by albitites. This is explained by the fact that ferrous and ferromagnesia rocks of the area lie in direct contact with a layer of sericite-quartz schists which also have undergone intensive metasomatism as indicated by their albitization. Evidently, these schists were the source of alumina in the albitization of aegirinites.

It is not impossible that in type-one columns, too, a portion of the albitite alumina was borrowed from sericite-quartz schists. This is especially so if we consider that at the present time it is difficult to draw a line between the two kinds of albitites. One criterion may be the presence of relict veins of crocidolite and hydrobiotite and the development of glaucophane and hydrobiotite. Another criterion is the common development of chlorite.

The aegirinite albitization front is commonly accompanied by hematite or carbonate fringes which correspond to the hematite-hydrobiotite zone of type-one columns.

Columns of type three are developed in sericite-quartz and garnet-sericite-quartz schists, i.e., in rock with a large content of Al_2O_3-SiO_2. They are marked by a simple structure and by the absence of an "alkaline" zone similar to the magnetite-alkaline-amphibole and the aegirine. The original rocks are albitized directly. If they contain ferromagnesian minerals, as is the case in garnet-sericite-quartz schists, then chlorite and, in places, hematite-chlorite fringes are formed, which may be regarded as analogues of the hematite-hydrobiotite zone. A distinctive feature of column three albitites is their lack of alkaline femic minerals. Chlorite and a small amount of fine-grained magnetite are usually developed in the paragenesis, along with albitite. The last zone of type three column is of quartz.

These three types of infiltrational-metasomatic columns represent the examples of pure alteration of rocks of a definite original composition. In so far as intermediate varieties and interbedded rocks of different

omposition do occur, this leads to corre-ponding mixed metasomatism. For instance, egirine-alkaline-amphibole varieties are fairly ommon.

In addition, infiltrational-metasomatic for-nations are also developed in rocks of other riginal compositions. However, they are less ommon and are developed on a much smaller cale. Thus, microclinization was observed in he metasomatism of mica-microcline-quartz ocks of a sericite-quartz layer. Mica-arbonate and quartz-mica-carbonate rocks give rise to a tremolite zone which is replaced by an albite or a quartz zone. The appearance of the albite zone is directly related to the nicaceous material content in the source rock. Such instances are significant because they appear locally in rocks enriched with apatite and to a lesser extent with zircon. Metasoma-tism leads to a transfer and concentration of P and Zr as apatite and malacon. This con-centration originates at the front of albitiza-tion or quartzitation, which confirms the low mobility of P and Zr in this metasomatic process.

On the basis of the above description of infiltrational-metasomatic formations, the following sequence of mobilities for the com-ponents of this process may be assumed:

$$K_2O,\ Na_2O,\ MgO,\ FeO\ \ Fe_2O_3,\ SiO_2,\ P_2O_5,\ TiO_2\ \ ZrO_2,\ Al_2O_3.$$

DIFFUSION-METASOMATIC FORMATIONS

As previously noted, diffusion-metasomatic formations are best developed in the core of the syncline. They have originated as a result of metasomatic reactions at contacts of dis-similar rocks such as quartzites and dolo-mites, dolomites, and Al-silicate schists, etc.

These metasomatites, too, have a zonal structure, although there are substantial dif-ferences between them and infiltrational-metasomatic formations. First, the chemical-mineralogic composition of reactive meta-somatites is considerably more dependent on that of the original rocks than is the case for infiltrational columns. All metasomatic min-erals originate at the expense of the contact-ing rock components, without any large regrouping of the latter. In contrast to infil-trational columns with their single direction, differential displacement of substance, diffu-sion-metasomatic formations exhibit diffusion of the components of the contact rocks.

A feature of reactive metasomatites is the lack of sharp boundaries between mineral associations of individual zones. Instead, there are gradual transitions from one mineral association to another. This is in direct con-tradiction to D.S. Zorzhinskiy's hypothesis of diffusion-metasomatic zonation and requires a theoretical explanation [6].

Where infiltrational columns are character-ized by structures of replacement of some metasomatic parageneses by others, reactive metasomatites exhibit widely developed poikilo-blastic and diablastic, commonly very-coarse structures (Fig. 6, 7). Obviously, they sug-gest a quiet crystallization under nearly equilibrium conditions resulting from a poor circulation of solutions.

Diffusion metasomatites lack the concen-tration of scattered and secondary components.

The general morphology of reactive meta-somatites is very bizarre and depends directly on the permeability of rocks and on tectonic elements. Favorable tectonic structures of original rocks, instead of being camouflaged by metasomatism, as is the case for infiltra-tional processes, seem to be enhanced by the developed metasomatic minerals.

Depending on the lithology of contacting rocks, a number of formations may be desig-nated, of which five are best developed and most important.

Metasomatites are most common at the contact of quartzites and dolomites (carbonate rocks high in dolomite). This group comprises the following parageneses of metasomatic min-erals which form zones on the whole parallel to the quartzite-dolomite contact: 1) dolomite, 2) dolomite-talc, 3) dolomite-tremolite, 4) quartz-dolomite-tremolite, 5) quartz-tremolite-diopside, 6) quartz-diopside, and 7) quartz.

The chemical process is extremely simple and obvious, consisting of the migration of silica into dolomites and of Ca and Mg into quartzites. However, a correlation of the composition of metasomatic minerals reveals that dolomite endocontact carries talc with the maximum SiO_2 content and the minimum con-tent of bivalent bases; on the other hand, diopside with least CaO and MgO and with most bivalent bases is formed in quartzite endocontact.

	CaO	MgO	SiO_2	H_2O
Diopside	2.59	18.5	55.6	–
Tremolite	13.8	24.6	55.8	2.8
Talc	–	31.7	63.5	4.8

The explanation of this appears to lie in the high carbon dioxide potential in dolomites and its rapid decrease in quartzites. A high

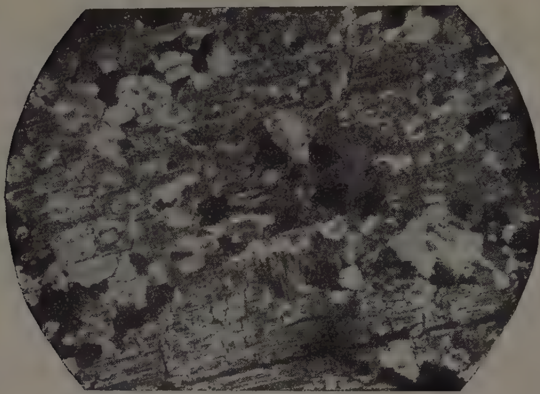

FIGURE 6. Poikiloblastic albite-diopside rock.
40X; crossed Nicols.

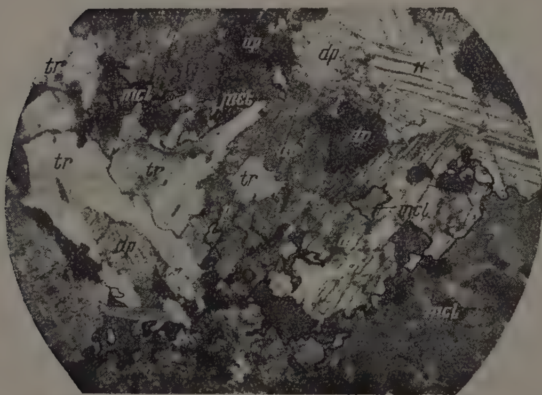

FIGURE 7. Diablastic structure in diopside (dp), tremolite (tr), and microcline (mkl).
10X; crossed Nicols.

CO_2 potential prevents the formation of silicates with a high Ca and Mg content. Instead, they are chiefly tied up in carbonates. A correlation of the compositions also suggests a higher mobility for Ca as compared with Mg.

The second fairly well-developed type, are reactive columns at the contact of quartzites with mica-carbonate and quartz-mica-carbonate rocks. The character of metasomatic changes here is very close to type one, with talc, tremolite, and diopside forming in the same sequence. The difference from type one is the wide development, in endocontact, of mica-carbonate albite rocks which occur with the same paragenesis as talc, tremolite, and plain carbonate. It is significant that the basicity of albite in carbonate rocks remains at the No. 4-6 level. This also points to a very high CO_2 potential. The mobility of alumina is extremely low, in this instance, inasmuch as albite is formed in metasomatically processed endocontact mica-carbonate rocks, nearly in situ, at the expense of micaceous material. Only in isolated instances of a considerable development of metasomatism, some migration of alumina into quartzites and the initiation of an albite-diopside paragenesis takes place (Fig. 6). As a rule, the appearance of albite is accompanied by the alkalinization of amphiboles. We have no chemical analyses for these amphiboles. In their optical properties, however, they are

closest to alkaline, calcium-carrying varieties of hastingsite and pargasite type [15].

Thus, in its most complete form, type two of reactive columns has the following composition: 1) dolomite-phlogopite (muscovite), 2) dolomite-talc-albite, 3) dolomite-pargasite-albite, 4) quartz-diopside-albite, 5) quartz-diopside, and 6) quartz.

We designate type three of the reactive-metasomatic formation to include columns at the contact of dolomites and magnetite-biotite-quartz schists. Besides SiO_2, CO_2, Mg, Ca, and Al, which participated in the formation of the first two types, iron plays an important part in the composition of the original rocks and their metasomatites. Iron determines the appearance of actinolite and its alkalized varieties of the ferrihastingsite type. Generalized, these columns have the following structure: 1) magnetite-biotite-quartz, 2) quartz-actinolite, 3) ferrohastingsite-albite, 4) dolomite-actinolite, 5) dolomite-tremolite, 6) dolomite-talc, and 7) dolomite.

It should be noted that the zone of development for ferrohastingsite (alkalized tremolite) is usually much wider than for albite, commonly embracing the dolomite endocontacts. The ferrohastingsite paragenesis carries fine-grained magnetite which forms poikilitic growths.

This succession points to a greater mobility of iron as compared with alumina, with iron usually migrating into dolomite. This is rarely observed for alumina (appearance of albite in dolomite endocontacts).

Of great interest are contacts of quartzites and dolomites with rocks high in potassium. Here belong quartz-biotite and microcline-biotite-quartz schists. A feature of these reactive columns, which we assign to types four and five, is the appearance of microcline. K-minerals are almost completely lacking in metasomatic formations of the area. Therefore, the microcline development can be explained only by a high-K content in the original rocks, assisted by a poor circulation of solutions.

Type four includes reactive columns along the contact of quartz-biotite and microcline-biotite-quartz schists, with quartzites. As a rule, quartzites and schists carry an admixture of carbonates with schists carrying some magnetite. The columns have the following structure: 1) microcline-biotite-quartz dolomite, 2) quartz-microcline-tremolite (or actinolite), 3) microcline-tremolite-diopside-quartz (Fig. 7). 4) tremolite-diopside-quartz, 5) diopside-quartz, and 6) quartz.

Microcline of zones two and three is an independent metasomatic generation with no direct connection with that of original rocks, appearing as it does in quartzite endocontacts. It forms coarse xenomorphic bodies, usually found in diablastic intergrowth with tremolite and diopside. It should be noted that there are formations intermediate between columns two and three, where albite is formed along with microcline.

Type five of reactive columns originates at the contacts of dolomites with quartz-biotites and microcline-biotite-quartz schists which also contain some magnetite. The main feature of this column type is the formation of epidote at the expense of biotite. The structure is as follows: 1) dolomite, 2) dolomite-tremolite, 3) dolomite-actinolite, 4) microcline-actinolite, 5) actinolite-epidote-microcline, 6) epidote-(microcline)-quartz, 7) (microcline)-biotite-quartz. (In parenthesis, microcline of original rock). It follows that calcium is the most mobile element whose occurrence in schists is marked by the formation of epidote which ties up, in addition, the Fe and Al of biotite. The appearance of Mg brings about the development of actinolite in the paragenesis with metasomatic microcline. The latter evidently requires for its formation, a high potassium concentration in the pore-filling solution. This seems to be achieved through the liberation of K from the biotite lattice. The mobility of alumina is not high, in this instance, since its diffusion is not observed.

Besides these five, there are other, less common reactive-metasomatic formations. As an example, there is the origin of aegirine at the contact of carbonate rocks with high-iron quartz-magnetite-biotite schists.

The above-named reactive formations suggest that the sequence of the mobility of components in a diffusion-metasomatic process is the same as in the infiltrational, namely:

$$K_2O,\ Na_2O,\ CaO\ \ MgO,\ FeO,\ Fe_2O_3,\ P2O_5,\ TiO_2,\ ZrO_2,\ SiO_2,\ Al_2O_3.$$

It was noted above that the poor filtration of solutions throughout lithologically homogeneous sections results only in their recrystallization or in the forming of some new minerals at the expense of in situ components. Therein lies the similarity of this process with metamorphism, as far as its physical-chemical nature is concerned [6]. Study of such phenomena is significant in relation to the composition of metasomatic solutions. In quartzites and dolomites, such a process results only in recrystallization. In mica schists, the formation of scattered albite is

observed. Tremolite or pargasite with albite originates in mica-quartz-carbonate rocks; and tremolite and diopside, in quartz-carbonate. It is not impossible that a portion of the cummingtonite schists originated in the same way from magnetite-biotite-quartz schists, with some of the aegirinites originating from ferrous quartzites.

SUMMARY

1. All metasomatic formations of the area originated as a result of a single hydrothermal-metasomatic process, in the framework of a single structural-tectonic plan.

2. All metasomatic minerals, including the accessory, originated from the components of source rocks, with their composition and relationship determined by the composition of these source rocks and by the differential mobility of components in the metasomatic process. Sodium is the only component brought in by hydrothermal solutions.

3. Both the infiltrational and diffusion-metasomatic formations have a zonal structure determined by the sequence of the components' mobility, which is the same in both instances, namely:

$$K_2O,\ Na_2O,\ CaO,\ MgO,\ FeO,\ Fe_2O_3,\ P_2O_5,\ TiO_2,\ ZrO_2,\ SiO_2,\ Al_2O_3.$$

4. It appears that the composition of all original hydrothermal solutions was carbonate-sodium, which would explain the formation of all the observed metasomatic minerals.

5. A concentration of secondary and scattered elements (TiO_2, P_2O_5, ZrO_2) is possible only in an infiltrational-metasomatic process with the presence of these components in the original rock as a prerequisite.

The author expresses his gratitude to I.V. Aleksandrov, A.I. Strygin, and A.I. Tugarinov for their close cooperation in this study.

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Received November 21, 1958

GEOLOGIC STRUCTURE STRUCTURE AND ORIGIN OF THE OGLANLY BENTONITIC CLAYS (TURKMENIAN S.S.R.)[1, 2]

by

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The industrial value of the Oglanly bentonites results from their exceptional adsorption, dispersion, swelling, and colloidal properties. The Oglanly deposit is a principal source of bentonitic clays extensively used in foundries, oil (adsorbents and drilling muds), porcelain, and other industries. Despite nearly 25 years of exploitation, its origin is not adequately known, which hampers to a considerable extent the discovery of new deposits of the Oglanly type in Central Asia.

The Oglanly bentonite deposit is located 36 km northwest of the Dzhebel Station, Ashkhabad Railroad, and 137 km east of the city of Krasnovodsk. It is associated with the left limb of the Bol'shoy Balkhan anticline which trends nearly due east-west for some 100 km and is made up of intensely dislocated deposits ranging from Upper Jurassic to Paleogene. These deposits, in a south-north sequence, form a series of parallel ridges trending from SE to NW; the largest is a ridge of dense Upper Jurassic and Neocomian limestones, sandy marls and sandstones, dipping steeply to the northwest and locally overturned (Danian-Senonian ridge). The overlying Tertiary rocks and bentonites are associated with a chain of mountain ridges generally known as the Kosha-Seyra, with individual summits, Kyariz, Oglanly, Gerkez, Karoyman, and others. Bentonitic clays are associated with Paleogene deposits at the very top of the local section; they are overturned and dip to the southwest, 60° to 80°. The bentonitic outcrops on the northern slope of the Tashli-Tepe ridge received most study. Its geologic map is given in Figure 1.

The base of the local section is made of dense, finely crystalline, slightly sandy limestones, as much as 500 m thick, light gray to white; they are very uniform, bedded, Senonian to Danian in age, and crop out in the southwestern part of the deposit.

They are overlain by lower Tertiary deposits which may be divided into the following formations, reading upward.

1. Kyarzin -- Pg_1^{kr}, represented by green to brick-red sandstones, with a total thickness of 115 to 125 m. Its contact with the underlying Danian limestones and calcareous sandstones is fairly well exposed east of the Kyarzin pass in the eastern area of the deposit. The Kyarzin formation changes upward to a formation of dense bentonitic marls and bentonitic clays (Oglanly formation). There is no apparent unconformity between these two formations, although their lithology is very much different.

2. Oglanly -- Pg_1^{ogl}, total thickness about 110 m. Its generalized structure is as follows:

a) gray bentonitic, thin-bedded marls, dense, locally sandy ("lower marls"), carrying intercalations of green plastic clays at their base, 20 m thick. This layer, underlying the bentonitic clays, varies somewhat in adjacent areas, with thin lentils of green to green-gray clays as much as 1.7 m thick locally in its upper part. In some localities of the western sector, their thickness decreases to 9 m.

b) light-gray, greenish-gray, locally brown bentonitic clays, 9.5 to 23.7 m thick.

c) thin-bedded bentonitic marls (upper marls), about 80 m thick; they, too, carry several layers of green clays and are light gray, green, or brown in color. They are platy, weathering to thin laminae with

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[1] Geologicheskoye stroyeniye i genezis oglanglinskogo mestorozhdeniya bentonitovykh glin (Turkmenskaya SSR).

[2] The author is deeply grateful to Doctor of Geologic and Mineralogic Sciences, V.P. Petrov, for a number of valuable suggestions; and to Candidate of Geologic and Mineralogic Sciences, B.P. Belikov and Doctor of Geologic and Mineralogic Sciences, M.F. Vikulova, for their study of bentonites with the electronic microscope.

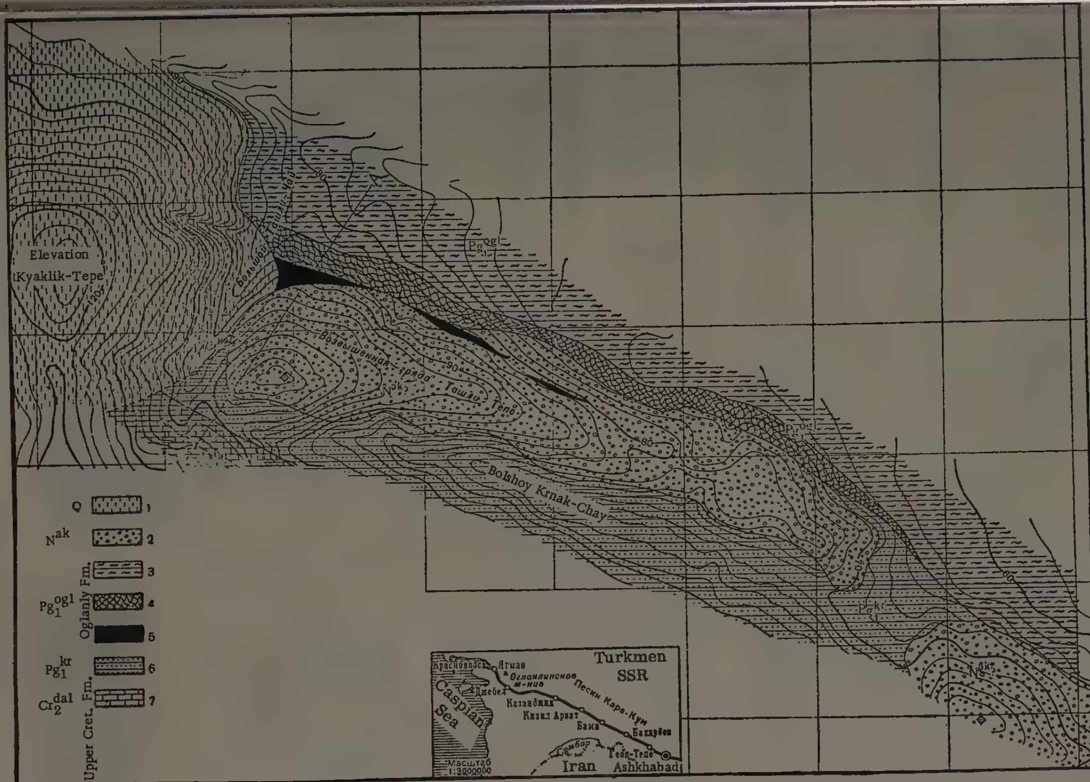

FIGURE 1. Geologic map of the Oglanly bentonite deposit (western sector). Compiled by S.P. Shobolov.

1 -- alluvial deposits, area of thick loesslike loams; 2 -- Nadakchagyl formation, a sequence of pebble beds and conglomerates; 3 -- upper, thin-bedded tripoli beds alternated with green clays (bentonitic marls of A.V. Danov); 4 -- bentonites; 5 -- lower, gray bentonitic marls underlying bentonites; 6 -- the Kyariz formation, gray, calcareous and argillaceous, also green sandstones; 7 -- Danian stage, light gray sandy and crystalline limestones.

The map is oriented by the magnetic meridian. Elevations from a conditional zero. Contour interval, 2 meters.

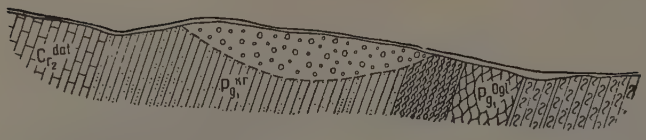

FIGURE 1a. Cross section along line A-A on geologic map.

surfaces usually strewn with fish scales and teeth. Well-preserved impressions of whole fish are observed, in places. An abundant fauna of foraminifera and radiolaria also has been found (Fig. 2). This foraminiferal assemblage is common in the Caucasian upper Eocene beds, whereas the character of marls, fish skeletons, and scales from the Oglanly formation suggests a correlation with the upper Eocene of Mangyshalk and Kara-Bogaz-Gol.

Microfauna in bentonites is represented by a comparatively small number of species but by a great number of individuals. This, most likely, suggests some peculiar conditions in the basin, probably related to the chemical composition of water.

The role of radiolaria in bentonites is very important, inasmuch as their siliceous tests may have been the source of silica in the Oglanly rocks.

A generalized cross section of the bentonitic deposit may be represented as follows (from north to south):

A. Western sector (Fig. 3).

1. Bentonitic marl, 0.7 m.

The underlying bentonites are subdivided by the author into two layers: upper and lower.

2. The upper layer is made up of bentonites, as follows:

1) green, plastic, with a brown cast; 0.10 m;

2) green, dense, hard, breaking under a blow into sharp angular chips; 0.8 m.

3) green, with conchoidal fracture, locally gypsiferous; 0.5 m;

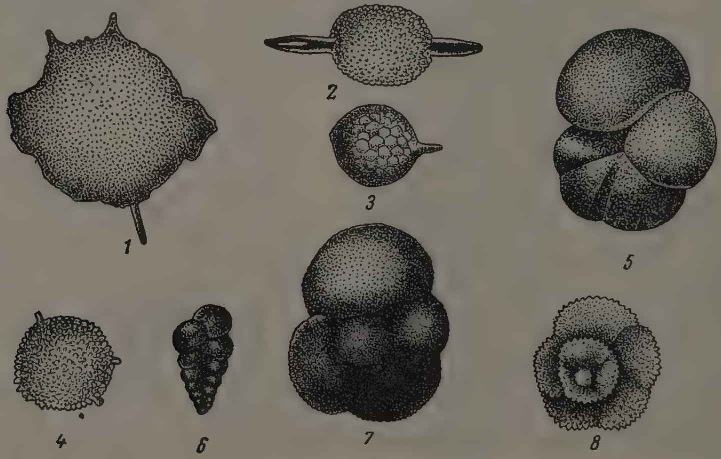

FIGURE 2. Microfauna of bentonites and bentonitic marls.

1-4 -- Radiolaria, 90X; 5-8 -- Foraminifera; 5 -- Nonion micrus Cole., 60X; 6 -- Gumbelina ex gr. globifera, 90X; 7 -- Globigerina ex gr. bulloides d'Orb.; 8 -- G. crassaformis Galloway et Wissler var., 60X.

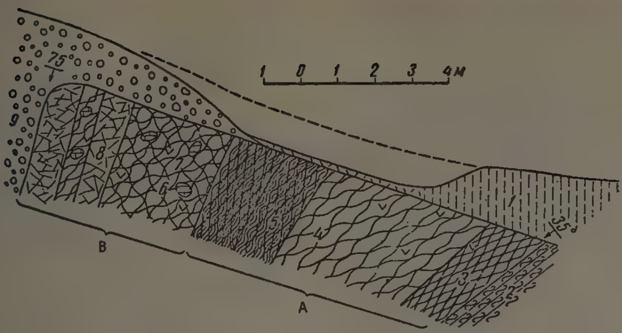

FIGURE 3. Cross section of quarry 7 (western sector).
A -- lower layer; B -- upper layer.

1 -- loesslike loam; 2 -- tripoli; 3 -- bentonite, green, rocklike; 4 -- bentonite, gray, waxy; 5 -- bentonite, green, friable; 6 -- calcareous-sandy concretions; 7 -- bentonite, gray, with conchoidal fracture; 8 -- bentonite, white; 9 -- pebble beds.

4) gray, waxy, dense to friable, with :her-colored admixtures, gypsum crystals, ıd biotite scales; 4.95 m;

5) green, friable, locally light green, ense, with biotite; 2.8 ;

6) green, foliated, very plastic ticky), typical soapstone; 0.3 m.

3. The lower layer is represented by:

1) bentonite, gray, dense, with con-hoidal fracture, fissured, with gypsum and .6 x 0.7 m concretions of white calcareous andy rock and lumps of light olive-green lay; 2.5 m;

2) bentonite, white, friable; 0.65 m;

3) bentonite, gray, waxy, with con-retions of white calcareous sandy rock and ypsum; 0.55 m;

4) bentonite, white, waxy, with small).15 x 0.3 m) concretions of calcareous-andy rock; 0.75 m;

5) bentonitic marl, light gray with a ypsum intercalation, 1 cm thick, and olive-ray clay, the so-called "shirt" layer at the ontact with pebble bed, dipping south at 75°; .10 m;

Thickness of the upper layer varies from to 9.5 m; the lower, 3.8 to 13.5 m.

In their texture, these bentonites are sub-ivided into three varieties: 1) gray, waxy, locally friable; 2) gray-white, rocklike; 3) dense, with conchoidal fracture, splitting under blow into sharp angular chips.

In the explored part of the western sector, the thickness of bentonite across the strike varies from 9.5 to 23.7 m. In borehole 2, at a depth of about 90 m, it is only 1.45 m thick. With the increase in calcareous material, bentonites change to the overlying and underlying marls, with sharp and distinct contacts between them.

B. Eastern sector.

This sector of the Oglanly deposit is located on the northern, uplifted part of the ridge which makes a sort of extension of the Tashli-Tepe ridge, cut by the Kyariz pass some 2 km east of the latter. A bed of light- to dark-gray bentonitic clay here is 1.45 to 9.2 m thick, an average of 5.5 m.

A facies change of bentonites to bentonite-like sandy clays is observed here; locally, bentonites are interbedded with marls. Bentonite is nearly devoid of calcareous concretions; gypsum occurs chiefly in a zone near the southern contact with the marls.

Composition of bentonite as a substance. The wide use of bentonitic clays in various branches of industry calls for a comprehensive study of their physical-chemical nature. Therefore, bentonite from Quarry 7 of the western sector was subjected to a detailed laboratory study by the author. It was tested for two main purposes: 1) for the foundry industry and 2) its use as an adsorbent.

After the war, the Oglanly bentonites were studied for use as an addition to porcelain batch and to drilling muds, as well as a plasticizer in calcinating materials.

Composition of bentonite as a mineral. A typical bentonite from a functioning quarry is represented by montmorillonite. Its texture is pelitic, with a marked orientation of thin scales forming a semblance of reticulate structure. There are floating grains, mostly small and angularly rounded, of feldspars (plagioclases), quartz, biotite, less commonly of zircon and nontransparent ore minerals. Locally, the rock is saturated with oxides of iron.

Bentonitic marl from pit No. 3 is made up of foraminifera tests cemented with a claylike mineral having a refraction index similar to that of montmorillonite, $\gamma = 1.517$. The shells are mostly leached and replaced by crystalline calcite.

Thus, montmorillonite is the principal clay mineral of the Oglanly bentonite where it forms sheaflike aggregates of drab-brown fibers. First among the impurities is quartz (as much as 10 percent of dustlike grains) followed by cristobalite, carbonate (chiefly in fine fractions imbedded as minute grains in the clay matrix), and small irregular grains of gypsum, feldspar (plagioclase), biotite and muscovite, limonite, and halloysite.

Chemical composition. Chemical analyses of typical bentonite varieties of the Oglanly and other deposits of bentonites and bentonitic marls are given in Table 1. The high SiO_2 content in the Oglanly bentonites appears to be determined by the presence of a considerable amount of amorphous silica (19 to 26 percent). The presence of CaO and CO_2 and the negligible content of SO_3 are explained by the presence of gypsum and carbonates.

Thermal- and electron-microscope study. Thermal study of the Oglanly and Azkamary bentonites have demonstrated their identity with the Askana bentonite (see Fig. 4).

Thermograms for all these bentonites, as well as for a typical montmorillonite, are characterized by three endothermic breaks: first, from 100° to 200°C, associated with heat absorption in the elimination of adsorped water; second, from 600° to 700°C, brought about by the elimination of water of constitution; and third, from 800° to 900°C, a result of the breaking down of the mineral lattices and of the amorphization of the substance.

The electron-microscope study has revealed that both the Oglanly bentonite and the underlying bentonitic marl carry isolated grains and botroidal aggregates of cristobalite in the

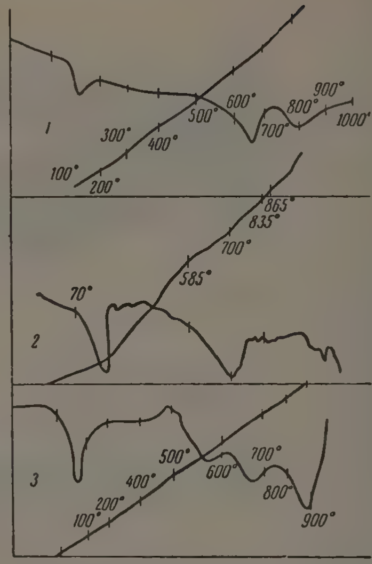

FIGURE 4. Thermograms for bentonites.

1 -- Oglanly; 2 -- Azkamary; 3 -- Askana.

montmorillonite matrix. This is confirmed by Debaye crystallograms taken by I. V. Iogansen [2].

Physical-chemical properties of bentonites. Principal industrially important properties of bentonites as components of foundry additives, porcelain batches, and drilling muds are their adhesiveness, bentonite number, gelation capacity, and degree of dispersion.

Adhesiveness is one of the decise factors in the avaluation of bentonites. It was determined by the McKean method, with the bentonite wetted to various extents, from 30 to 70 percent, at 10 percent intervals. Adhesiveness of the Oglanly bentonite, with 30 to 40 percent moisture, was about 1 kg/cm^2. The highest adhesiveness, 1.68 kg/cm^2 at 30 percent moisture, was exhibited by a number of Oglanly bentonite samples. The Askana bentonite gave a 0.25 adhesiveness with 60 percent moisture, and the maximum of 0.72 kg/cm^2 with 35 percent moisture. The American bentonites show an adhesiveness of 0.72 kg/cm^2 with 40 percent moisture. The adhesiveness of different clay from the Soviet Union is 0.3 to 0.4 kg/cm^2, at 25 percent moisture.

Thus, the Oglanly bentonites possess an exceptional adhesive capacity exceeding that

Chemical Analyses of the Oglanly and other Bentonites

Nos	Locality	Rock	SiO_2	TiO_2	Al_2O_3	Fe_2O_3	CaO	MgO	K_2O	Na_2O	SO_3	Cl	CO_2	H_2O >100°	H_2O <100°	Total
1	Oglanly, Quary 7 Specimen 48	Bentonite, gray, waxy.	71,19	0,20	13,75	1,54	1,42	2,72	1,52	2,26	0,19	0,80	3,02	3,1	8,99	100,3
2	Same, specimen 52	Bentonite, green, sticky, foliated	64,21	0,20	11,13	2,37	1,32	2,49	1,85	2,42	0,18	0,43	7,72	0,51	8,13	100,29
3	Same, specimen 53	Marl at top of bed	64,53	0,30	2,73	3,02	10,81	1,03	1,11	0,90	0,34	Not det.	Not det.	10,86	3,07	—
4	Oglanly, East sector	Bentonite, gray	62,20	Not det.	12,10	—	2,94	2,70	—	2,12	0,14	0,07	2,94	15,88	7,80	—
5	Azkamary, pit 2, bed 2	Bentonite, gray	53,70	—	17,18	5,37	0,66	3,20	0,50	2,92	0,36	Not det.	0,44	15,42	8,44	99,90
6	Same, bed 3	Bentonite, gray	54,30	—	16,65	4,40	0,84	3,20	0,70	3,23	0,51	»	0,40	15,38	8,68	99,61

Note: Comma represents decimal point.

of the known Askana (askangel) and American samples.

An important characteristic of bentonites is their bentonite number, determined by the V.A. Aronovich method, as follows: 20 g of clay is placed in a vessel with 500 c.c. of distilled water. The mixture is stirred up for one hour by an agitating machine and left to rest for 24 hours. After that it is decanted and weighed, and the weight of the remaining water is computed in percent of the original. The bentonitic number of common bentonites is more than 50 percent; and of high-quality bentonites, as much as 95 percent. Sandy, noncolloidal clays have a low bentonitic number, not more than 10 percent.

Of the tested Oglanly bentonites, 70 percent gave a bentonitic number of 80 to 98 percent, with only a few more than 50 percent. On the whole, after a 24-hour settling period, the Oglanly bentonites give a stable colloidal solution.

It may be mentioned for comparison that the bentonitic number of the Askana bentonites varies from 60 to 95 percent, and the American "clayspar" from Wyomong, from 70 to 100 percent.

Bentonite is characterized by rapid swelling upon wetting, with the sorption increasing the volume by a factor of 10 to 15 and with the formation of a gellike mass. According to N.K. Sazonova, the Oglanly bentonite swells 18 times its original volume. The diameter of bentonite particles is 0.20 to 0.35 micron.

The plasticity and the binding capacity of bentonites are especially important in the manufacture of porcelain. The plasticity of the Oglandy bentonite, after Attenberg, is 36.8, with 18.5 for kaolin. This figure marks the Oglanly bentonite as a high-quality, grade-one plastic clay.

Because of its high plasticity and binding capacity, the Oglanly bentonite is used as an addition to drilling muds, along with more fusible low-grade clays. Muds prepared with bentonite clays possess high structural and thixotropic properties, maximum stability, low specific gravity, and low water loss. An addition of bentonite (2 to 5 percent of volume) considerably improves drilling muds. Coarse particles of more fusible clay apparently form a framework filled with finely dispersed bentonite particles.

Activated Oglanly bentonites may be used as bleaching agents in the refining of petroleum products (cracking benzine, kerosene, lubricating oils), especially transformer oil, vaseline, and vegetable oil.

In special cases, colloid properties of the Oglanly bentonitic clays may be activated by processing the clays with sodium fluoride (6% NaF).

* * *

Two types of bentonitic clay deposits are recognized, both in this country and abroad:

1. Deposits of a definitely volcanic nature, made up of volcanic rocks, ash, and volcanic glass.

Such are most of the Caucasian deposits: the Askana with its alkaline andesite tuffs, the Gumbrian and Nal'chik; abroad, Tertiary deposits in Wyoming, the Dakotas and Arizona (U.S.A.); bentonites of the island of Ponza (Italy) and Ichtenmessee (Hungary), the latter carrying rhyolite tuffs; also the Manghatu and Poranghau deposits (Howke Bay, New Zealand).

2. Bentonite deposits whose volcanic origin has not been directly established.

In the Soviet Union, the Azkamary bentonites (Bokhara province, Uz. S.S.R.) occur resting among Paleogene calcareous clays (Suzak-Khanabad stage), and the mid-Volga deposits (Alekseyevskoye, 40 to 45 km from Kuybyshev, and others); abroad, Eocene bentonitic clays of the Paris basin and Miocene continental clays of North Africa (Morocco).

The Oglanly bentonites belong to type one. Proof of their volcanic nature is found in thin section where pseudomorphs of montmorillonite on volcanic glass and associated cristobalite may be observed. The marine conditions of the accumulation of ash as a matrix rock are suggested by a marine microfauna in the clays.

The above data raise the question as to the source of the volcanic ash. In their most recent work, A. A. Ali-Zade and M A. Rotko (1) describe a widespread development of volcanic ash in various layers of southwestern Turkmenistan, such as the red beds, the Akchagylian, Apsheronian, and Bakuian stages. N. I. Andrusov associates the Akchagylian ash on Cheleken Island with submarine volcanic eruptions of the Pliocene.

On the other hand, considering the possibility of the long-distance transfer of ash and the lack of major contemporaneous volcanic structures, plus the findings in Azerbaydzhan of tuffs with a fairly similar composition, a transfer of ash material from the Caucasia centers may be surmised.

SUMMARY

1. The Oglanly bentonite ("oglanlite") is an alkaline montmorillonitic, high grade, finely dispersed clay.

2. Electron-microxcope and X-ray study demonstrated the presence of cristobalite in bentonite, which is explained by the volcanic nature of the clay source material.

3. The large amount of radiolaria and foraminifera, together with the lateral changes of bentonite to marl, suggest that the deposition and decomposition of volcanic ash took place in a marine environment.

4. Intercalations of volcanic ash in southwestern Turkmenia may constitute evidence of bentonitic clays. Especially promising are Paleogene deposits west of the western sector of the Oglanly deposit.

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Received May 28, 1958.

BRIEF COMMUNICATIONS

FIRST DISCOVERY OF FOSSIL REPTILES IN THE TUNGUSSKA BASIN[1]

by

B.P. V'yushkov and A.I. Yemel'yanova

During the 1956 geologic survey in the Tungusska basin, the participants in expedition No. 5 of the Aerophotogeology Trust discovered remains of fossil reptiles in beds between the Permian and Triassic. They were found in the lower course of the Korvunchana River (more correct, Korbunchan), a left tributary of the Nizhnyaya Tungusska River some 300 km from the mouth of the latter, in the Evenkiysk national district of Krasnoyarsk province.

The following section has been measured by A.I. Yemel'yanova on the right bank of the Korvunchana, 2.5 to 3 km from its mouth:

P_2^2 (?) 1. Sooty and carbonaceous shales with thin sandy limestones and calcareous siltstones. The latter contain poorly preserved pelecypod (?) shells. Apparent thickness, as much as 4 m.

T_1^1 (?) 2. Gray to green and blue-gray, highly calcareous tuffaceous sandstones and siltstones; 1.0 to 1.5 m thick.

3. Greenish-gray to gray-green dense tuffaceous sandstones abundant in spherical and ellipsoid formations made up of dense dark-gray siltstones, locally carrying reptile bones. The latter were observed 1.5 m above the water mark. Siltstone pebbles locally occur in the sandstones. Thickness about 5 m.

4. Tuffaceous sandstones and siltstones, gray-green, with some dolerite pebbles and chunks of tuffs: unidentifiable plant remains and pelecypod shells, Ustchamiella tungussica Rag., U. opinata Rag., U. obrutschevi Rag., and others (identified by L.A. Ragozin). Thickness, 3 to 4 m.

5. Thin to thick flagstones of fine- to coarse-grained tuffaceous sandstones and siltstones, with poorly preserved plant remains. Apparent thickness, 3 to 5 m.

The exposed beds are flex into very gentle folds with dips of 5° to 10°.

A similar section is observed on the left bank of the Korvunchana, three km from its mouth. Its distinguishing feature is the presence of dense siltstones with conchoidal fractures. M.N. Blagoveshchenskiy collected abundant remains of Cladophlebis sp. from these beds.

The collection, transferred to the Paleontologic Institute of the Academy of Sciences, U.S.S.R. (No. PIN 1312) contain the following items: a neural arc, probably of the sixth or seventh lumbar vertebra, a fragment of sacrum, the distal head of a large tibia, a a metacarpal bone, and fragments of ribs and unidentifiable bones. All of the remains, judging from their size and state of preservation, belonged to the same individual, namely to a comparatively large representative of the order Dicynodontia. A more or less complete skeleton seems to have been buried at this site.

In their size and general aspect, these remains are close to dicynodont remains at Gor'kiy II site (Lagernyy Ovrag), assigned to the upper half of the Tatarian stage. The poorly ossified tibia head without the helical joint, distinguishes this animal from Rhadiodromus klimovi Efr. from the Donguzsk site (Sol'-Iletsk region, Chkalovsk province, Tananyk formation, middle layers of the Lower Triassic). This difference means that the Tungusska specimen was more primitive and therefore more ancient. The upper end of the neural arc is reminiscent of that of lumbar vertebrae belonging to Lystrosaurus weidenreichi Young, from Lower Triassic deposits of Sin'tszyan. The poorly preserved tibia head is also suggestive of genus Lystrosaurus. All of this fixes the age of the remains as the uppermost Permian or, what is more probable, the base of the Triassic. Additional data are needed for a more definite correlation.

The position of the bone-bearing layer in the section is very noteworthy. The very base of the section in this area is marked by

[1] Melkiye soobshcheniya.

coal-bearing deposits with thin limestones (bed one, above) assigned to the top of the Permian. They are overlain by the so-called Terrigenous sequence of thin, not more than 25 m, fine-grained sandstones and siltstones with some tuffaceous material. This sequence is not present everywhere, but rather occurs in patches. Besides the Korvunchana section, where it carries our find, its presence is established for the Chambakon valley, along Degal' and Uchama Rivers and along the lower course of Bol'shaya Bugarikta. The latter locality is noted by the brown and drab color of its rocks. The age of the terrigenous sequence is determined as the basal Triassic, from the Dicyonodont remains, from pelecypods which L. A. Ragozin believes to be reminiscent of the Balbino fauns (Jurassic), and from Cladophlebis remains. According to M.F. Neyburg, these are unlike the typical Jurassic forms of Siberia, but are undoubtedly younger than Permian. Their Lower Triassic age is established also by V.D. Prinada who described Cladophlebis kirjamkensis Pr., C. Pectophora Pr., C. jeniseica, etc.

Higher up in the section, there lies the so-called tuffaceous sequence whose thickness does not exceed 100 m. Even a cursory analysis of this section reveals that the terminal Permian sedimentation took place in a depressed marshy area subject to occasional sea invasions. At the Permian-Triassic boundary, a general, although slow, uplift took place, which led to the transformation of this marshy land into steppe or forest steppe. Volcanic activity began at the same time, culminating in the formation of the tuffaceous sequence.

Viewed in this light, the association of the dicyonodont remains with the Tungusska basin terrigenous sequence is quite reasonable. Indeed, these bulky creatures, the denizens of savanna in most cases, and capable of long marches, could not prosper in marshy Permian lowlands. They penetrated there only after the land had been uplifted and had become a savanna. There are reasons to anticipate the discovery of large dicynodont sites in Triassic deposits of the Tungusska basin. On the whole, the Tungusska find undoubtedly is of great paleostratigraphic and paleogeographic interest.

In the summer of 1957, the Korvunchana site was studied by members of the Nizhnetungussk party of the Paleontologic Institute at the Academy of Sciences, U.S.S.R. It was established that it was the burial site of a single large dicynodont. It is located at the lower end of the first large right-bank exposure, 2.3 km from the mouth of the Korvunchana River. The bones rested in a green-gray hackly siltstone, 2 m above its contact with dark carbonaceous siltstones at the top of the Upper Permian (Degalinsk formation). The skeleton was considerably damaged by recent erosion. Only a few vertebrae, some limb bones, and bone fragments were recovered. A bone fragment was also found in the nearest up-stream exposure, on the right bank.

The overall thickness of light gray, hackly to thick-slated siltstones, tuffaceous siltstones, and fine-grained sandstones of the terrigenous sequence, exposed in the three large right-bank exposures and in one on the left bank, is tentatively estimated at 25 to 30 m. They are overlain by dark to blue-gray tuffs and tuff breccias of the tuffaceous sequence.

Significantly, the right-bank contact between these two layers is very sharp and distinct, whereas it is lacking as such in the left-bank exposure where the transition is achieved by the appearance of tuff concretions and lenses at the top of the terrigenous sequence. They rapidly grow in number, and finally take the place of siltstone. This, together with the better development of tuff breccias, suggests that the deposition took place in the immediate vicinity of an eruptive center. The exposed rocks along the Korvunchana are cut by several pyroclastic dikes and are microfolded.

Besides the dicynodont remains, associates of the Paleontologic Institute have collected, from the Korvunchane-exposed section, a large number of fern-leaf imprints and other fossils, from green siltstones of the left bank. Dark argillites of the Degalinsk formation (Upper Permian; bed one, above), exposed on the right bank, 4.5 km from the mouth, yielded a large number of ostracods, phyllopods, and some fish remains. The collected material is being processed and identified.

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Received May 3, 1957

NEW DATA ON THE AGE OF THE ARMASU FORMATION, NORTHERN TIAN-SHAN[1]

by

V.S. Burtman and V.Ya. Medvedev

In the middle thirties, V.G. Mukhin separated the Aramsinsk formation of volcanic-sedimentary rocks, in the basin of the

[1] Novyye dannyye o vozraste aramsinskoy svity severnogo Tyan'-shanya.

Susamyr River. Similar rocks were described subsequently by V.N. Kozerenko, from the southern slope of the Kirghiz range [1]. Because of the lack of fauna, the question of the age of these formations remained moot. V.G. Mukhin assigned the Aramsinsk formation to the upper Paleozoic. V.N. Kozerenko dates as Devonian-lower Carboniferous the extrusive clastics of the Karakol and Aramsu Rivers. V.A. Nikolayev placed this sequence at the base of the lower Carboniferous red beds. The latter view has been adopted for all small-scale geologic maps of the area.

At the present time, a wide development of these deposits has been established for the Kirghiz and Susamyr ranges and for the eastern part of the Talas Ala-Tau. An Ordovician fauna was collected from the Aramsinsk formation. It has been established that only the upper part of the volcanic-sedimentary section, given by V.N. Kozerenko for the Kirghiz range, is correlative with the Aramsinsk formation. Older deposits, separated from the latter by a minor break, are designated as the Barkol' formation. The overlap of Aramsinsk rocks on the upper Barkol' beds is well established in the Susamyr (Aramsinsk basin) and Kirghiz ranges. Extrusives of the Barkol' formation are intermediate to basic; those of the Aramsinsk formation, intermediate to acid.

Volcanic-sedimentary deposits rest with an angular unconformity upon Middle Ordovician and older rocks. They are cut by granitoid intrusions and are overlain unconformably by lower Visean conglomerates. They differ from the underlying formations in the low intensity of their deformations.

The Barkol' formation is developed in the

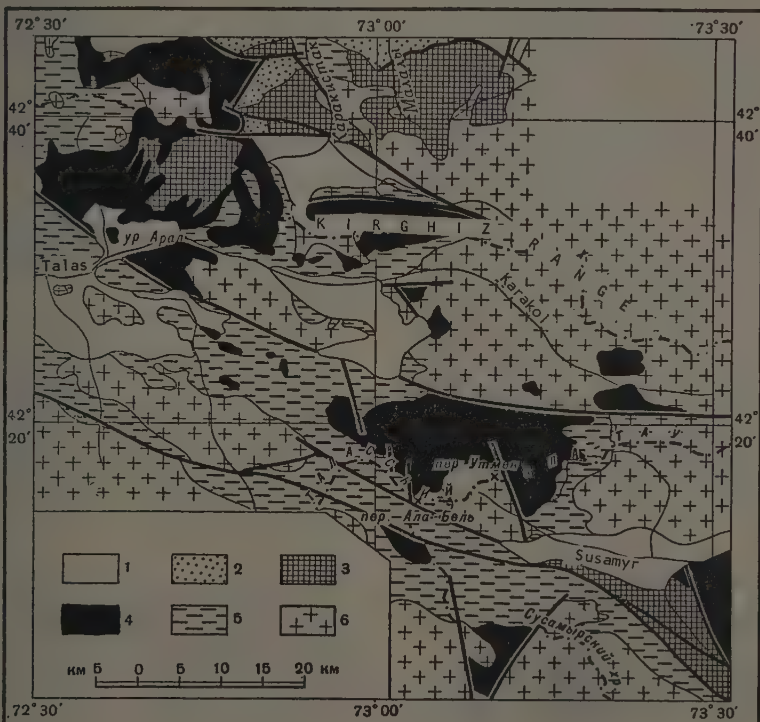

FIGURE 1. Distribution of Upper Ordovician deposits.

1 -- Cenozoic deposits; 2 -- Lower Carboniferous deposits. Upper Ordovician deposits: 3 -- Aramsinsk formation; 4 -- Barkol' formation; 5 -- pre-Upper ordovician deposits; 6 -- intrusions.

Kirghiz, Talas, and Susamyr ranges. Its type section is located in the Barkol' River basin, north of the settlement of Aral (northern slope of the Kirghiz range). It is made up of porphyrites, tuffs, polymictic sandstones and conglomerates (with pebbles of porphyrites, sandstones, siltstones, and less commonly granites). The rocks are green to purplish-red. Their thickness reaches 600 m.

Rocks of the Aramsinsk formation are developed in the Susamyr and Kirghiz ranges. In the latter, the formations are nearly all extrusive, whereas tuffaceous rocks predominate in the former.

The most complete section is described from the northern slope of the Susamyr range. Here, in the Aramsinsk basin, the Barkol' porphyrites are overlain by 300 m of lilac-colored conglomerates, gravels, sandstones, and acid extrusives. Best developed in conglomerate pebbles are granites (35%), green polymictic sandstones and siltstones (30%), and acid extrusives (10%). Significantly, porphyrites which underlie the conglomerates are missing in the pebbles. Higher up in the sections, there are 450 m of alternating lilac-red to green tuffaceous gravels and conglomerates with porphyrites. The section is capped by about 1000 m of lilac-red tuffaceous, commonly cross-bedded, gravels with wedging-out layers of acid extrusives, chiefly quartz porphyries. Beds are 10 to 25 m thick.

A bed of dark-green siltstones lie in the upper part of the Aramsinsk formation, 300 m from the top. It carries numerous small brachiopods (collections of D. M. Shteyman, V. S. Burtman, and I. P. Pugacheva, 1956).

Among them, E. N. Yanov identified Lingula aff. subcrassa Eichw. and Lingula sp n. The first form, according to E. N. Yanov, "corresponds to L. subcrassa from orthoceratites Ordovician limestones of the Baltic region, differing from the latter in its small size"; the second is "similar to Lingula orbicularis Eichw. from the Ordovician of the Baltic region, Salair, and Kirghizia but is smaller and carries radial grooves." N. V. Litvinovich identified from the same collections, the species Philhedra laelia Hall, common in the Upper Ordovician of America.

The lower age limit of the volcanic-sedimentary sequence is established definitely. Rocks of the Barkol' formation rest with an angular unconformity upon a sandy siltstone sequence where A. V. Grigor'yev found, in 1955 (in the Keptasha River basin) brachiopods Dinorthis sp., plectambonites sp. (identified by O. I. Andreyeva), and orthoceratites Proterocameroceras sp. (identified by G. Z. Balashov). In the same year, V. I. Knauf collected from the same sequence along the Dzhaysan River, Rhynochotrema cf. otarica, Strophomena sp., Ampyz sp., akin to A. tenuispinosus; Harpes sp., Pliomera sp. (all identified by T. F. Rukavishnikova); also Rukavishnikova); also graptolites, somewhat higher up in the section: Dicellograptus sp., Diplograptus sp., Climacograptus sp. (identified by A. M. Obut). On the basis of this fauna, deposits underlying the Barkol' formation may be assigned to the Llandiloian, and possibly to the lower Upper Ordovician.

The Barkol' and Aramsinsk volcanic formations rest unconformably upon Middle Ordovician deposits. They carry an Ordovician fauna in their upper part. Accordingly, we assign them to the Upper Ordovician. The question of Silurian elements in them cannot be answered at the present time because of the the lack of data: paleontologically defined Silurian deposits are unknown in northern Tyan'-Shan.

The establishing of an Upper Ordovician age for volcanic deposits affords a new interpretation of the history of this area at the close of the lower Paleozoic, as follows:

1. Upper Ordovician deposits rest unconformably on the intensively dislocated Cambrian and Lower and Middle Ordovician rocks which are folded, as a rule, into large brachysynclines. Thus the main Caledonian stage of folding is associated with the onset of Upper Ordovician. After the Proterozoic, this is the most intensive folding of the area.

2. The incidence of vast granitoid intrusions in this area does not coincide with the main Caledonian folding stage. It takes place in the middle Paleozoic.

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All-Union Aerogeological Trust
Ministry of Geology
and Protection of Deposits of the U. S. S. R.
Moscow

Received May 22, 1957

STRATIGRAPHIC DIFFERENTIATION OF THE DONBAS LOWER CARBONIFEROUS, BY MICROFAUNA[1]

by

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A study of the microfauna of this section permitted the subdivision of lower Carboniferous deposits of the western part of the Donbas into 32 members (local).

They were traced nearly everywhere (areas of Staro-Beshevsk, Kurakhovo-Mar'inskoye, Petropavlovskoye, and Mezhevoy-Volch'ya) and partially correlated with contemporaneous sediments of the Dnieper-Don and Galitsiysk-Volnsk troughs and the Moscow basin.

The proposed correlation of the lower Carboniferous in the Donbas is based on numerous data from boreholes and exposures in the above-named localities. The area of study embraces approximately 5,000 square kilometers. More than 7000 limestone samples have been studied.

Given below is a brief description of the lower Carboniferous by groups and individual members.

Lower division of Carboniferous C_1, Tournaisian stage C_1^t. Tournaisian deposits are subdivided into four members (I, II, III, and IV), approximately corresponding to zones $C_1^{ta,b,c,d}$ of A.P. Rotay, or partially to faunal beds XVI and XV of N. Ye. Brazhnikova, separated in the area of the Dnieper-Don trough. These deposits differ in their foraminiferal content and in the limestone facies. At the base, limestones are usually fine grained, with sparse ostracods, to barren dolomite; followed by finely crystalline limestone, chiefly algal; and finally by argillaceous limestone with abundant ostracods. Algae (tubular) of the algal limestones are widely developed in Tournaisian and lower Visean deposits.

An abrupt faunal and lithologic change occurred at the Tournaisian-lower Visean boundary. Argillaceous limestones with abundant ostracods and primitive small foraminifera (Parathurammina, Archaesphaera), which form the top of the Tournaisian, changed at the base of the Visean to crystalline foraminiferal limestones abundant in Quasiendothyra, Endothyra, and tubular algae. This difference in the faunal assemblage suggests a difference in sedimentary conditions of the time.

Members V, VI, and VII are recognized in the lower Visean of this section. They are rich in foraminiferal fauna and are correlative with zones C_1^{va-e} of A.P. Rotay and to microfaunal beds XIV and XIII of N. Ye. Brazhnikova, established for the Dnieper-Don trough. Quasiendothyra staffeliformis is abundant at the base of this interval, followed by Haplophagmella, Lituotubella, Permodiscus, Forschiella, etc. The uppermost of the three, member VII, differs from the others by its intercalations full of sponge spicules. Marine fauna (brachiopods, foraminifera, corals) are very rare. In their foraminiferal fauna, these members are very similar to zones (Oleskov, Bus, and Yakhtorov) designated by P. L. Shulga for the Galitsiysk-Volynsk trough, also to the upper part of the coal measures and the Tula member of the Moscow basin. Deposits of C_1^vf of A.P. Rotay (corresponding to N. Ye. Brazhnikova's microfaunal bed XII-a and the base of XII) are also subdivided into three members, VIII, IX, and X, with foraminiferal assemblages considerably different from each other. Calcifolium algae, widely distributed in the upper part of the Visean, appear for the first time at the base of the lowest of these members. The rich and diversified foraminiferal assemblage with Saccamminopsis and typical Mikhaylovo forms noted in foraminiferal and algal detrital limestones provides a basis for correlating the overlying members with the Vladimirov and Ustiluzhsk zones of the Galitsiysk-Volynsk trough and the Mikhaylovo beds of the Moscow basin.

Deposits resting between the top of zone C_1^{vf} and limestone bed V_7 (in the Artemuglegeologiya Trust nomenclature, and represented by detrital foraminiferal and algal (Calcifolium) limestones, also contain a small amount of fully grown Mikhaylovo-type fauna among which begin to appear, in ever increasing numbers, dwarfed Archaeodiscus krestovnikovi Raus., Endothyra bradyi Mikh., and other forms. This interval, which we subdivided into members XI, XII, XII, is closest in its foraminiferal fauna to the upper part of microfaunal bed XI of N. Ye. Brazhnikova, the lower part of the Poritsa zone of the Galitsinysk-Volynsk trough, and to the Venevsk member of the Moscow basin. Thus, the interval from the top of zone C_1^{ve} of the Donbas to the limy bed V_7 finds its correlatives in the Oka formation of the Moscow basin, of the same sedimentary cycle.

Stratigraphically higher, a marked

[1] Stratigraficheskoye raschleneniye nizhnego karbona donetskogo basseyna po mikrofaune.

impoverishment in fauna is observed in the upper part of zone C_1^{vg} of A.P. Rotay (calcareous beds V_8-C_5, in the Artemuglegeologiya Trust nomenclature), commonly in crinoidal and detrital limestones with dolomitized intercalations. Some typical large Mikhaylovo forms disappear (Archaediscus moelleri var. gigas Raus., Endothyra omphalota Raus. et Reitl., Bradyina rotula (Eichw.) etc.), whereas Archaediscus krestovnikovi Raus., Quasiendothyra ukrainica Brazhn., Endothyra bradyi Mikh. begin to appear (at times in mass). There is a sharp decrease in the number of Calcifolium algae.

Six members are separated in this interval: XIV, XV, SVI, XVII, XVIII, and XIX, which we correlate by foraminiferal fauna with foraminiferal beds X, XI-a and the lower part of IX, of N. Ye. Brazhnikova; with the Poritsa (upper part) and Ivanichi zones of the Galitsiysk-Volynsk trough, also the Tarussk and Steshevsk members of the Moscow basin.

A new stage in the development of foraminiferal fauna is initiated beginning with limestone beds C_5-C_6 (base of zone C_1^{na} of A.P. Rotay) which are characterized by the abundance of crinoids and brachiopods. Here appear for the first time such forms as Eostaffella protvae Raus., Eostaffella paraprotvae Raus., Archaediscus(?) namuriensis Dain.

First appearance of Bradyina cribrostomata Raus. et Reitl. and large Eostaffella postproikensis Vdov. is noted somewhat higher up, in detrital limestones with foraminifera and Calcifolium algae (C_7-C_8 in the Artemuglegeologiya Trust nomenclature). Along with the younger fauna, this interval (upper part of the Donbas coal measures, or C^{vn} of the unified system, or C_1^{na} of A.P. Rotay) contains typical Visean forms, Monotaxis, Hyperammina etc., also Calcifolium algae. The latter have not been found higher up in the Donbas section. This interval is divided into four members: XX, XXI, XXII, and XXIII. The cited foraminiferal assemblage of the upper part of the Donbas coal measures, which we assigned to the lower Namurian, is typical for sediments correlative with microfaunal beds IX (upper part) and VIII of N. Ye. Brazhnikova, also to the lower part of the Lishnyany zone from the Galicia-Volyn' trough and the Protvino member of the Moscow basin.

The overlying sequence (formations C_1^4 of the Geologic Committee [GeolKom], or C_1^{nb-d} of A.P. Rotay, with the exception of limestone bed D_1, which gravitates faunally to our lower Namurian -- is divided into four members, XXIV, XXV, XXVI, and XXVII.

The lowest bed carries a sparse mixed-type foraminiferal fauna. Along with the younger, newly appeared forms (Globivalvulina, Ammodiscus compactus), more ancient, superannuated forms are present (Eostaffella mediocris, Endothyra crassa). They are widely developed in Visean deposits. Corals are abundant, especially in the area of the Kal'mius River. The two following members, with an impoverished foraminiferal fauna, are marked by the appearance of oolites,[1] algae Donezella litugini Masl., and hydroactinids, which are also developed higher up, in sediments of the Baskhirian stage. The fourth member, with an even poorer foraminiferal assemblage, is marked by an alternation of detrital, Donezella, and oolitic limestones. Typical Visean forms are utterly lacking. This interval (formation C_1^4 of Geolkom), in its lithology and foraminiferal fauna, belongs to the upper part of the Namurian.

Formation C_1^5 of Geolkom is divided into three members, XXVIII, XXIX, and XXX. The lower member is represented by sandy detrital to detrital-Donezella limestones with sea anemones, a dwarfed foraminiferal fauna (Archaediscus ex gr. baschkiricus, Eostaffella pseudostruvei, E. varvariensis, etc.).

The middle member, represented mostly by limestones with algae (Donezella), sea anemones, and stromatopora (?) intercalations, is marked by first appearance of Pseudostaffella antiqua Dutk. and by an increasing number of small Eostaffella varvariensis, archaediscs etc. The upper member, represented essentially by the same types of limestone (except for the stromatopora ?) contains more of the small eostaffels and archaeodiscs.

We assigned these deposits (formation C_1^5 of Geolkom), characterized by a middle Carboniferous aspect fauna, to the base of the middle Carboniferous Bashkirian stage.

Two members are separated within formation C_2^1 of Geolkom. Member XXXI (beds F_1, F_1^1) is represented mainly by fine-grained limestones, poor in fauna, typical (in their lower part) of relatively deep-water deposits.

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[1] Oolitic limestones are widespread in the eastern part of the subject area (Kal'mius basin) and are scarce in its western parts (Kurakhovo-Mar'inskiy, Mezhevaya-Volch'ya). Oolites have not been found in the Petropavlovsk area of the eastern extension of Bol'shoy Donbas.

Member XXXII (beds F_1^2, F_1^3, F_1^4, F_2P is represented by algal (Donezella) to detrital-algal, locally hydroactinoid limestones with fairly common foraminifera whose assemblage is enriched with the appearance, from the bottom up, of such new species as Bradyina nautiliformis Moell., Ozawainella umbonata Brazhn. et Pot., Pseudostaffella antiqua (Dutk.) var. grandis Schlyk., Novella manukalovae Brazhn., etc.

The above-named foraminifera are also common in the overlying Bashkirian middle Carboniferous rocks.

Artemuglegeologiya Trust,
Artemovsk, Stalino Oblast

Received December 19, 1957

CENTENNIAL OF MICROSCOPIC PETROGRAPHY[1]

by

A. S. Marfunin

Beginning of micropetrography. In 1858, an Englishman, G. Sorby published his work "On Microscopic Structure of Crystals," dealing with the origin of minerals and rocks. Following the tradition set up by F. Yu. Loewinson-Lessing half a century ago, we accept that event as the beginning of the microscopic method in petrography.

That work was preceded by a number of other remarkable discoveries of a more general character, without which its writing would have been impossible. First, there was the advent of crystallooptics. E. Malus' discovery of the polarization of light, in 1808, modified the concept of light which had prevailed for more than two thousand years, since the time of Plato and Euclid. In one or two decades after that, the brilliant work of A. Fresnel and D. Brewster discovered all the basic laws and defined the basic concepts of crystallooptics, which are still valid with but a few refinements. We are also justified in noting at the sesquicentennial of optical crystallography, the discipline which has played such an important part in all sciences dealing with crystalline bodies, and especially in petrography.

In 1828, W. Nicol obtained polarized light vibrating in a single plane, by sawing a rhombohedron of Iceland spar and glueing the halves together. Six years after, G. Talbot built the first polarizing microscope with an analyzer. W. Nicol was the first to prepare a thin section of petrified wood; in 1851, Oschats demonstrated a whole collection of thin sections of minerals and rocks, such as labradorite, leucite, Chalcedony, almandite, perthite, marble, crystals in glass, etc. In 1850, Sorby described a thin section of calcareous sandstone.

It was not until a hundred years ago, however, that microscopic study was applied to determining the origin of rocks and minerals, thus opening great possibilities to students. In this connection, it is important to emphasize that petrography, since its inception, has not been only a descriptive discipline.

In the sixties of the last century, pioneer works in micropetrography appeared in many countries, among them papers in Russia by A. Inostrantsev (1867?) and A.P. Karpinskiy (1869).

The following decade witnessed the appearance of fundamental works by F. Zirkel (1873), H. Rosenbusch, F. Fouquet, and A. Michel-Levi, which contained the basic descriptive body of petrography and were instrumental in disseminating the microscopic method.

Initial steps of optical mineralogy. In one of his early works on the application of the microscope, the famous Czech mineralogist and petrographer, H. Cermak emphasized that "the great progress of microscopic analysis of rocks is due to the fact it was not limited to the observation of mineral forms under the microscope but made use of polarized light in determining the structure and optical properties of transparent minerals in order to obtain reliable criteria for the major mineral forms."

Optical constants of rock-forming minerals were determined by A. des Cloizeaux, beginning in 1855, first without the microscope, on cleavage plates set between crossed nicols; then with polarizing devices, with the application of parallel and convergent light (by 1864, he described all kinds of dispersion of optical axes) and by means of the improved microscopes of Amichi and Neremberg employing a rotating table.

The study of thin sections required a technique different from that designed for determining minerals from fine grains and from various cross sections. In other words, the rock-forming minerals had to be re-identified in thin sections. Such diagnostic features were first established by F. Zirkel for feldspars,

[1] Sto let mikroskopicheskoy petrografii.

nepheline, leucite; by H. Cermak for amphiboles, pyroxenes, biotite, etc. (extinction angles, pleochroism, conoscopic figure).

Almost simultaneously, A. Lazo, E. Bertrand, and K. Klein combined the microscope with the conoscope. In 1878, it acquired a form not essentially different from what it is now.

However, it measured in a single plane, only. As a result, optical constants were determined only in very precisely oriented, thin sections of sufficiently large crystals or else in definite cross sections of minerals in thin sections of rocks.

In the first case, various polarizing devices were used, such as those measuring the angle of optical axes with a precision within several minutes. In determining the minerals from thin sections, on the other hand, a most painstaking method was worked out for the determination of the angle of extinction in oriented cross sections, normal to optical axes, or to the sharp and obtuse bisectrices, and determined either conoscopically or in sections oriented relative to cleavage. The use of half-tone devices and the stauroscope made it possible to establish the position of extinction to a fine precision.

In 1881, M. Schuster constructed a curve of plagioclase extinctions as a function of its anorthite content. This was first (and not only in petrography) "composition-optical properties" diagram which marked the beginning of a new trend in the study of optical properties of rock-forming minerals.

Initiation of the immersion method. Early determinations of the refraction indices for minerals were made with refractometers. But even Sorby had sought a way to determine the refraction of minerals from their thin sections. O Maschke (1872) used oblique light for correlating the refraction indices, a method which gained wide recognition after the work of J. L. K. Schroeder van der Kolk (1900). Most significant was F. Becke's discovery (1893) of the light streak ("Becke line") on the boundary between two media of different refraction, which made the correlation of their refraction indices very simple. F. Becke successfully used this method in the identification of plagioclases. The "Becke line," along with the effect of oblique light, initiated the immersion method of determining the refraction indices.

The Fedorov method and a new stage in the development of petrography. Following the microscopic, chemical, and physical chemical stages in the history of petrography, a new period began in connection with the unprecedented development in geologic surveying, prospecting, and exploration, which called for a wide application of petrographic and related microscopic study.

This period of greater demand on and for petrographic descriptions brought about a wide application of the Fedorov method of microscopic study of minerals. Its acceptance as a practical means of petrographic mensuration was slow in coming. The original model of the universal table was offered by Ye. S. Fedorov in 1891. It required 20 to 30 years for it to become an essential tool of petrography, first in this country and then abroad.

The Fedorov method is the second birth of micropetrography. It not only affords a means of observation of rocks in thin sections and of the identification, with a desirable degree of precision, of their component minerals, but also of a rapid and precise measuring of their constants, thereby determining the ratio of these components in a mineral.

The principle of both the conoscope and the Fedorov method is the same: it is the study of interference phenomena for a crystal in various directions.

In the first method, this is achieved by means of convergent light which gives a general picture of the interference phenomena for a given beam; the second affords a study of these phenomena for each direction, separately. The first method may be called synthetic; the second, analytic.

The Fedorov method is preferable in the study of thin sections.

Significantly, the introduction of the Fedorov method led to a wide application of the composition-optical properties curves. Thus, the first edition of the well-known A. N. Winchell's manual of optical mineralogy (1909) contains only two such curves; the second edition (1927), 39; the third (1933), 56; and the fourth (1951) more than 129 (see preface to the fourth edition).

In recent years, the application of this method made it possible to outstrip the immediate potentialities of microscopy, by affording an insight into submicroscopic phenomena -- the processes of structural regulation in minerals.

Development of microscopy during the last few decades. The mensuration technique of the optical constants of minerals has been intensively developed in the last 20 to 30 years, through the discovery of new theoretical possibilities and through the charting of the arduous course from general considerations to the specific conditions of measuring

various mineral groups from thin sections. Only a few principal discoveries and improvements can be mentioned here.

In the field of the determination of mineral refractions, there is the interference microrefractometer (A.N. Zavaritskiy), the rotating needle method (A.G. Kolotushkina), the double immersion method (R.Ch. Emmons, N.M. Melankholin), the prism method as applied to small crystals (M.K. Bel'shterli), a combination of the immersion and Fedorov methods (I.A. Ostrovskiy), the method of focal screening (Yu.A. Cherkasov), the application of high-refringence phosphorous and other immersion liquids. The immersion method as applied to thin sections (N.I. Nakovnik) and to ground rocks (V.P. Petrov) also has been worked out, along with a method of total determination of optical properties of minerals in immersion liquids (V.B. Tatarskiy).

A calcite rotating compensator was proposed for measuring the difference in path (V.V. Nikitin - M. Berek), using one elliptic compensator for small differences in path, and another for very large differences (A. Eringhaus). A method of birefringence determination in a conoscope is described by L. Borgstrem.

Especially numerous are the works dealing with the Fedorov method. Here belong a detailed method of measuring on a pentaxial table (R.Ch. Emmons, A.N. Zavaritskiy), the zonal method of plagioclase determination (A. Rittman), the determination of the extinction angle for monoclinic pyroxenes and amphiboles (D.S. Korzhinskiy), the stereoscopic method of determination of small angles between optical axes (L.A. Vardanyants), a combination of the Fedorov and conoscope methods (V.V. Nikitin, A.N. Zavaritskiy, V.A. Zavaritskiy, G. Schuman), the method of determining the orientation of optical inticatrix from the extinction curves (I.D. Muir), the triad method of determining the laws of twinning in minerals (L.A. Vardanyants), the construction and application of inclined tables -- hemispheres for the Fedorov method (V.V. Arshinov), the use of the table as a refractometer (N.M. Melankholin, V.N. Lodochnikov), the refractor microgoniometer (A.N. Zavaritskiy), and the construction of a measuring table for high temperatures (E. Wood). A method of quantitative description of mineral coloring has been perfected (N. Ye. Vedeneyeva, S.V. Grum-Grzhimaylo, N.M. Melankholin) along with the description of a method for the rapid identification of rare earths in rocks, by means of a spectroscopic ocular (S.V. Grum-Grzhimaylo). A phase-contrast attachment for microscopic observations has been invented (F. Zernike).

Various auxiliary graphs and nomograms have been devised to facilitate computations: N.K. Razumovskiy'a cyclogram; nomograms for determining the principal refraction indices, from measurements in a given section (I.A. Ostrovskiy, A.A. Lyapunov, M.V. Pentkovskiy); for the same purpose in any section (Z. Koritnig); graphs for determining the angle between optical axes in conoscope, and of differences in path in V.V. Nikitin—M. Berek's compensator (V.P. Petrov); the application of Woolf's grid to the solution of crystallooptical problems has been considered, on the basis of the relation between the optical orientation and the extinction angles in various sections, the angles of optic axes, the refraction indices in any section, etc. (V.A. Nikolayev, G. Terch); the application of vector analysis to crystallooptical problems (C. Burry); a method of expression of optical orientation by means of Euler's angles, and its application in computation (C. Burry) etc.

Interpretation of optical properties of minerals. Four stages may be designated in the development of our concepts on the importance of optical properties in the identification of minerals.

1) it was originally believed (A. des Cloizeaux, F. Fouquet, A. Michel-Levi, and others) that each mineral is characterized by definite optical constants.

2) After G. Cermak had demonstrated that feldspars present a continuous series of three components, M. Schuster presented in his diagram for plagioclase extinction angles, the continuous change in optical properties with corresponding change in composition. This principle of continuity and correspondence, subsequently presented in a general form by N.S. Kurnakov, has been made a basis for numerous composition- optical properties diagrams. Their most recent review is found in a paper by V. Treger.

3) Numerous measurements, performed by petrographers, have shown considerable variations in optical constants, irrespective of the composition. D.S. Belyankin demonstrated the geologic significance of these variations in the example of anorthoclases from younger intrusions in the Caucasus. A generalization of such observations in measuring plagioclases has led A. Keller (1941) to the differentiation between the "high" and "low temperature" optics for plagioclases. O.F. Tuttle has arrived at the concept of two contrasting mineralogies for extrusive and intrusive rocks. The relationship between optical constants of minerals and the latter's geologic position was considered by V.P. Petrov.

4) V.S. Sobolev (1949) was first to conceive the possibility of optical properties of plagioclases being dependent on the formation of

"superstructures" (patterns of solid solutions). X-ray study has established the nature of transformations in feldspars as a function of Al and Si.

From these data, we have subdivided the optical constants of minerals into the more and less structurally sensitive components, and built the first diagram in coordinates ("composition against optical-orientation-pattern") for plagioclase.

Of great importance in the interpretation of optical properties is their consideration from the crystallochemical point of view (V.S. Sobolev).

A promising beginning also appears in the direction of the determination of selective relations of individual optical constants for this or that mineral group with their content of various components, the construction of individual diagrams for minerals whose content varies in intrusive and extrusive rocsk (M.M. Veselovskaya), and even of the relation between the change in optical properties of minerals and the change in their composition within a massif or an ore deposit (A.A. Marakushev).

Place and future of microscopy among petrographic methods. Modern optical instruments, the methods of measuring and interpretation of optical constants, and their promise to petrography are a result of more than a hundred years of continuous development and improvement. Although there is an urgent need for new microscopic methods, and for the improvement of present methods of rock study, (in X-ray analysis, isotope analysis, study of magnetic, mechanical, and other properties of rocks) it may be stated, nevertheless, that microscopy is the principal method of laboratory study in petrography. The advent of new concepts, theories, and other methods of study only enhances rather than lowers its value. It stands in the same relation to a subject of study as the length of X-rays to inter-atom distances in crystalline substances.

The possibility of an absolutely perfect method of mensuration is limited to a certain extent, but it is very tangible as far as applied petrography is concerned. Specifically, there is the urgent task of raising the precision of the Fedorov table in the determination of the composition and structure of minerals from their constants. The method of conoscopy on the Fedorov table is very promising in this respect.

The highway of progress in micropetrography lies through the vast possibilities of interpretation of observations and measurements.

We now notice in thin sections much of what went unnoticed by earlier petrographers; and there are many more observations to be made. Petrographic research differs from a petrographic determination in its search for the new.

The guiding idea, now, behind such observations and measurements is paragenetic analysis which is a logical step in the development of theoretical petrography.

REVIEWS AND DISCUSSIONS

"ENGLISH-RUSSIAN GEOLOGICAL DICTIONARY" [1, 2]

by

M.V. Kalinko

The development of science, its broadening and deepening, calls for new concepts and new terms. It is quite natural, therefore, that specialized as well as general dictionaries are necessary in the study of foreign scientific literature. No general dictionary can embrace the entire range of ideas and terms pertaining to a discipline or a technology. In addition, a term commonly has different meanings in different fields of knowledge. For this reason, the compilation of a comprehensive dictionary, such as the "English-Russian Polytechnichal Dictionary," is not justified. In addition, the unavoidable bulk of such a dictionary hampers its use.

In 1937, Yu.S. Dyushen (Duchesne) published the first Soviet "English-Russian Geological Dictionary." This dictionary, for the most part, has become obsolete in these 20 years and is now a collector's item.

The publication of a new "English-Russian Geological Dictionary," by T.A. Sofiano fills the resulting gap and is very timely. The new dictionary compares favorably with the preceding one in the number of terms (about 26,000 words as against 14,600 in the Dyushen dictionary) and in their comprehensive treatment (see, for instance, such words as hade, notch, tilt, etc.).

In a very short time, the new dictionary has gained popularity and has become a handbook for all geologists interested in the foreign literature. Its edition (18,000 as against 4,000 for the Dyushen book) was quickly sold out, and a second edition is urgently needed.

[1] Ob "Anglo-Russkom geologicheskom slovare."

[2] English-Russian Geological Dictionary, compiled by T.A. Sofiano. Moscow. State Technical-Theoretical Literature Press, 1957.

In this connection, it is pertinent to point out certain shortcomings of the new book. First, a number of English terms are given a rather incorrect and too narrow an interpretation which, unfortunately, leads to erroneous translations. Thus, the word, "shale," is usually translated in dictionaries as "glinistyy slanets." However, it more often means argillaceous rocks in general; the adjective, "shaly," is more correctly translated as "glinistyy" (clayey, argillaceous) than "slantsevatyy" (bedded, platy), as "shale limestone." It is also important to point out that the term, "shale," is often used for rocks with porosity as much as 30 percent. Its translation as "glinistyy slanets" results in paradoxes such as "The source material of oil is deposited in schists (slantsy)" or "Finally, the time comes when oozes turn to schists" (A.I. Levorsen, Geology of Oil, Transl. from English, Gostoptekchizdat, 1958, pp. 388-389).

The term, "sandstone," indeed means "peschanik" (sandstone) in most cases, as given in the dictionary. Quite commonly, however -- and this is not mentioned by T.A. Sofiano -- it is used in a broader sense to mean sandy rocks in general. By the same token, the term, "limestone," refers to all carbonate rocks.

Not included in the dictionary are such composite terms common in geologic literature as "bottom structure," "depositional feature," "core salt," "mud logging," "oil shows," "lithological succession," etc.

The erroneous translation of the term, "silt," occurs in all dictionaries. Its earlier meaning indeed was "il" (slime, ooze), "gryaz'" (mud), "shlam" (sludge), etc. Now, however, it definitely designates clastic rocks with particles from 0.0039 to 0.0625 mm (1/256-1/16) or 0.005 to 0.05 mm in size. Their cemented variety, "siltstone," is "alevrolit," given as "alevrit" in the dictionary.

A reason for these erroneous translations is the evolution of individual words in the evolution of a language. One should not, therefore, mechanically transfer the translation of words from old dictionaries. Their meaning should be reviewed in the new context, as reflected in the latest literature.

One of the most glaring shortcomings of the new dictionary is the nearly total lack of a number of basic terms common to the related disciplines of paleontology, geophysics, geochemistry, paleobotany. Indeed, any geologic treatise nearly always considers the results of paleontologic, geochemical, geophysical, and other studies; these branches are so closely interlocked that there is no definite boundary between them. Therefore, a familiarity with the basic vocabulary of allied disciplines is essential in translating geologic literature.

It should be noted, in conclusion, that despite these shortcomings, the new dictionary is unquestionably superior to the old one. T. A. Sofiano has done much to help the geologist in his utilization of foreign literature.

It is to be hoped that both the author and the press will note these remarks in the publishing of the second edition which, it is taken for granted will be done. The "English-Russian Geological Dictionary" should be made more complete to come up to the present level of science.

CHRONICLE

ACCOUNT OF THE SECOND CONFERENCE ON GEOLOGY AND ORIGIN OF FERROSILICEOUS FORMATIONS OF THE UKRAINE[1]

by

Ya. N. Belevtsev

This conference took place at Krivoy Rog, on April 21-25, 1958, at the initiative of the Academy of Sciences U.S.S.R., and the Main Geologic Administration of the Ukr.S.S.R. Branch of the Ministry of Geology and Conservation of Mineral Resources, U.S.S.R.

Two hundred and fifty representatives of the above institutions participated, along with members of other scientific and industrial geologic organizations.

The conference opened with papers by N.P. Semenenko, Academician, Ukr.S.S.R. Academy of Sciences, on "Ferrosiliceous formations, their composition and position in the middle part of the Ukrainian crystalline shield:" and by Ya.N. Belevtsev, Corresponding Member, Academy of Sciences Ukr.S.S.R., on "The results of the study of the geology and origin of the Krivoy Rog ores." Thirty-six papers on regional problems were heard.

1. Geologic structure of the Krivoy Rog-Kremenchug ore belt was the subject of papers by M.N. Dobrokhotov (Kremenchug area), N.I. Polovko (Right-bank region), A.P. Korshenbaum (the Lenin and Red Guard mines); V.M. Maksimovich (the XX Party Congress and Frunze mines), V.M. Zaruba (the October Revolution and "Bol'shevik" mines), G.V. Tokhtuyev (the Liebknecht and Kirov mines), B.I. Goroshnikov (southern part of the Saksagan belt), V.Yu. Fomenko (southern limb of the main Krivoy Rog syncline), M.I. Chernovskiy (Tarapak-Likhmanovsk anticline), and R.I. Siroshtan (Inguletsk area). A detailed stratigraphic column for the Saksagan area was presented by S.A. Skudirin; stratigraphy of the upper formation, by D.I. Ishchenko and G.M. Struyev; a paper on magmatism of the Krivoy Rog basin was given by M.P. Kuleshov.

The papers presented the latest concepts on the complex folded and faulted structure of the Krivoy Rog ore belt and its component areas, all based on the latest drilling and prospecting data.

Of great interest were reports on new stratigraphic data in the detailed differentiation of iron ore formations into subformations and beds, also on the age relationship of the Krivoy Rog series, Saksagan plagioclases, and Zhitomir microcline-plagioclase granites. The recent studies by M.P. Kuleshov, B.I. Goroshnikov, A.I. Stratygin, and others disclosed and analyzed pebbles of plagioclase, microcline-plagioclase granites, ferrous hornstones, amphibolites, and even magnetite ores and arkosic conglomerates, at the base of the Krivoy Rog series. Mapping has revealed a northwestern trend (discordant with the Krivoy Rog trend) of structures in the plagioclase granite massif. These and some other data suggest an ancient, pre-Krivoy Rog age for plagiogranites, and the presence of ferrous hornstones, amphibolites, and even iron ores of a pre-Krivoy Rog age.

2. The origin of the Krivoy Rog type iron ores was dealt with in papers by Ya. N. Belevtsev, A.I. Cherednichenko, S.M. Ryabokon', Yu.P. Mel'nik, Yu.M. Yepatko, G.G. Bur, and A.I. Strygin. They were followed by a discussion, with Yu.G. Gershoyg, V.S. Fedorchenko, S.M. Zhilkinskiy, N.P. Grechishnikov, and many others, participating.

The latest studies, whose results were reported at the conference, have demonstrated the complex and multistage genesis of the iron-rich Krivoy Rog ores. The iron accumulation took place under sedimentary, metamorphic, and hypergene conditions. Sedimentation and diagenesis of ferrosiliceous deposits was the foundation of all ferrous rocks, and locally of rich iron ores, as it was established for the lower subformation of the upper formation. The second important stage of the iron accumulation in deposits was a period of dynamic thermal metamorphism connected with the first and main tectonic stages of the

[1]Itogi vtorogo soveshchaniya po geologii i genezisu zhelezisto-kremnistykh formatsiy ukrainy.

Krivoy Rog basin.

The third stage of formation and transformation of ores was hypergenesis and associated deep zones of oxidation of ferrous rocks and rich iron ores. The hypergenesis resulted in a considerable displacement of Fe, SiO_2 and other components, which brought about the change of magnetite ores to soft martite ores and the appearance of goethite-hematite and martite ores in schists and in iron-rich pseudospilites.

The following genetic ore types are designated in the Krivoy Rog basin, on the basis of the latest data: metamorphic, metamorphosed in hypergenesis, and hypergene. Thus, the long controversy on a hypergene versus a hypogene origin of the Krivoy Rog iron ores has been resolved to a considerable extent: both the metamorphic and hypergene processes played an unquestionably important part in the ore making.

3. Geologic structure and the patterns in the formation of ferrosiliceous rocks of the Ukraine were the subjects of papers by G.A. Makukhina (Verkhovtsevsk area), V.D. Ladiyeva (Konksk area), G.V. Zhukov (West-Azov region), V.F. Khallo (Belozersk area), and N.P. Semenenko (patterns in the formation of sedimentary-volcanic pseudospilites).

4. Hydrogeology of the Krivoy Rog basin was dealt with in papers by V.D. Natarov, K.A. Parkhomenko, and F.A. Rudenko.

5. Geologic structure and origin or iron ores of the Kursk magnetic anomalies (K.M.A.) were discussed in papers by M.N. Dobrokhotov (general patterns of development of the K.M.A. geologic structure), A.A. Glagolev (alkaline metasomatism in K.M.A. rocks), A.S. Yegorov (main types of disturbances and ore-controlling structures in the K.M.A. pre-Cambrian), I.A. Rusinovich and N.A. Plaksenko.

M.N. Dobrokhotov presented the K.M.A. geologic structure as consisting of two stages: the ancient, represented by gneisses and migmatites, similar to the Teterev-Bug series of the Ukraine or the Svionian series of the Kola Peninsula; and the younger metamorphic series of metabasic rocks, ferrous hornstones, and schists, along with carbonaceous-carbonate-argillaceous rocks. This second structural stage is quite similar to the Krivoy Rog series, being also divisible into three definite formations.

The majority of geologists, familiar with the K.M.A. deposits, believe in a hypergene origin for the ores of the Yakovlevskoye, Gastishchevskoye, Mikhaylovskoye, and Korobkovskoye deposits.

Judging from the data, however inadequate, a complex origin may be assumed for the K.M.A. deposits, with the iron accumulation achieved, as in the Krivoy Rog basin, through sedimentary, hypogene-metasomatic, and hypergene processes.

One of the papers was by S.M. Meleshkin, who represented the chairman of the Dniepropetrovsk Sovnarkhoz. He spoke on the future of iron mining in the Ukr.S.S.R., for the current Seven-Year Plan.

The conference recommended a further study of iron ore deposits and ferrosiliceous formations of the Ukr.S.S.R. Its most important objectives were set forth by the conference, as follows:

1. A study of concentration patterns of iron and other elements in the area of development of ferrosiliceous formations of the Bol'shoy Krivoy Rog.

2. A clarification of the deep structure of the Krivoy Rog-Kremenchug ore belt, and the change in ore deposits at depth.

3. A further study of the origin or iron ores in relation to the effect of hypogene and hypergene processes on the formation of various deposits.

4. Working out of prospecting methods for iron-rich ores of the Krivoy Rog type, under the conditions of magnetic anomalies' regions; and of the geophysical method of subsurface prospecting.

5. Hydrogeologic mapping of the Bol'shoy Krivoy Rog area, on scales of 1:50,000 and 1:25,000, in 1959 to 1963.

The Conference recommended annual meetings on the geology of ferrosiliceous formations of the Ukraine and Kursk magnetic anomalies.